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REFERENCE

Lane, B. H., 1935, Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C., 35c.)

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MIDDLE ORDOVICIAN BEDS IN THE SAGUENAY VALLEY, QUEBEC

G. WINSTON SINCLAIR

ABSTRACT. Beds of Middle Ordovician age have long been known in the Saguenay Valley of Quebec, but their age has been debated. Re-examination, in the light of modern work on equivalent beds further south, permits identification of horizons in the Black River and Trenton groups of the Mohawkian.

INTRODUCTION

THE presence of outliers of Ordovician rocks well within the margin of the Canadian Shield near Lake St. John has been known since 1829, when Baddeley described the area (1829, p. 125 ff.). These outliers were studied by Richardson (1858), and his collections were described by Billings and by Lambe. The area adjacent to Lake St. John was studied by Dresser (1916) and a second area north of Chicoutimi, some fifty miles to the east, was the subject of a report by Denis (1933), the paleontology being done by McGerrigle. The last two reports are readily available, and it will be assumed in these notes that the reader can refer to the detailed maps contained in them.

The age of the beds comprising these outcrops has been debated. Billings referred his fossils to the Black River or Trenton. Dresser quoted Raymond's determination of his fossils as indicating a horizon "either in the Black River or at the very base of the Trenton, more probably the latter." Bassler (1915), whose influence in matters of age assignment has probably been stronger than that of other workers, cited the Lake St. John fossils in his Index as "Black River or Richmond." McGerrigle (Denis, 1933) decided his collections indicated the age as Cobourg. All these workers assumed that the limestones were all of one age.

In view of these varied age determinations, there has been a natural reluctance to accept the presence of abundant *Halysites* low in the Middle Ordovician as proven. The known occurrence of beds of Richmond age on islands in Lake St. John has caused an understandable suspicion in the minds of workers elsewhere that some error might be involved.

In 1945 I spent four weeks in the vicinity of Roberval and Chicoutimi, in the employ of the Geological Survey of Canada, engaged in reconnaissance stratigraphy and collecting. Although this work left many problems unsolved, and many gaps which an instrumental survey will fill in, it did permit determination of the succession of the Middle Ordovician rocks of the outliers, and collection of sufficient material to permit tentative correlations and to serve as a basis for critical work by others. On a later trip to Chicoutimi I had the privilege of showing Dr. R. H. Flower over the section, and had the benefit of his opinions.

I wish to express my gratitude to the Geological Survey of Canada, and to Dr. W. A. Bell, then chief of the Palaeontological Section, for the opportunity of seeing this area. I must also thank Dr. W. K. Gummer, then of the Aluminum Company of Canada at Arvida, for his kindness in showing me outcrops on the Company's property, and giving me information based on test holes drilled by them.

Although my first trip to the district was made for the Geological Survey of Canada, and the material collected forms part of their collections, this paper does not reflect any official views.

In general the determinations of fossils are to be taken as first approximations, i.e. they have the value they usually do in faunal lists. The task of critical revision of the Trenton faunas is under way, but of necessity it is a slow one. Papers revising some small genera have appeared, three of them (*Dimeropyge*, *Eobronteus* and *Glyptocystites*) involving species from this area. Dr. R. H. Flower has undertaken study of the cephalopods and has published some preliminary notes on them (1952). The nature of these lists is stressed because of my conviction that the stratigraphy of the Trenton will not make sense as long as our fossils are referred to a limited number of "standard" species. So long as the bewildering variety of calymenids are all listed as *Flexicalymene senaria*, so long the calymenids will be worthless in correlation, and the same is

true of other genera where specific names are used so conventionally that they are useless or, worse, misleading.

Although I have suggested correlation of these beds with parts of the New York and Ontario sequence, I feel that use of formational names defined in these distant areas would be misleading at this stage of our study. Accordingly, I have introduced local names to permit unambiguous reference to the beds of these outliers. If later work shows the Tremblay to be indeed the local expression of the Lowville, then the local name can be dropped. For the present its use emphasizes that more study is needed for complete understanding of the relationships. The name Simard had been used informally by Dresser and McGerrigle for the limestones as a whole in the Simard area. I here use it formally, and apply it to a part of the sequence. I consider correlation between Chichoutimi and Roberval to be sufficiently certain to permit use of the same terms in both regions, although, as I have indicated below, there are some beds which I cannot place with certainty.

OUTLIERS IN THE SIMARD AREA

Denis and McGerrigle have described the areal distribution of the patches of limestone and shale which are found north of Ste. Anne de Chicoutimi. They treated the limestones as one unit, and on that basis McGerrigle suggested a Cobourg age for the whole. Since their work was done new quarries have been opened, and new data on the thickness of beds have been provided by exploratory drilling by the Aluminum Company of Canada, so that a reinterpretation of the section can be attempted.

There are two groups of outcrops, one about three miles northwest of Ste. Anne, which we may call for brevity the Tremblay area, the other at Chute aux Galets on the Shipshaw river, some twelve miles further northwest. Denis and McGerrigle are probably correct in postulating a connection between the two areas, but no lucid outcrops were seen in the country between them, and they may be distinct.

Tremblay area.—Throughout the area of outcrop indicated on Denis' map, small patches of limestone are met in the bush, but seldom are more than scraps of beds seen. Solution along joints has produced a karst topography, on a very small scale. Two quarries expose partial sections, and three small strippings

along the road show a few feet of beds. These would be unintelligible were they not tied in by the knowledge that they occur 58, 70 and 100 feet above the Precambrian, as determined by drill records. From these strippings, and the exposures in the quarry of the Quebec Bureau of Roads, the following section can be restored:

7.	Thick-bedded, somewhat sugary limestone, with rather thick, shaly partings. (stripping no. 3) (Simard?)	4 ft.
	Covered	22
	Simard Beds	
6.	Thick-bedded, dense, blue-gray limestone, weathering yellowish. No chert or silicification; barite rather common. Only one specimen of *Halysites* seen (stripping no. 2)	8 ft. 6 in.
	Covered	10
5.	Thick-bedded limestone, essentially one bed with an underlying shaly seam exposed. Small (4 in.) *Halysites* and *Hesperorthis* common. (stripping no. 1)	2
	Covered	18
4.	Rather thin-bedded limestone with silicified fossils and some plates of chert. Large heads (10 in.) of *Halysites* common; many actinoceroids and other cephalopods	10
3.	Dense gray limestone, weathering bluish, in beds up to 18 in. *Halysites* present but rare, *Streptalasma* common. Some bedding planes covered with indeterminable smooth trilobites	15
	Tremblay Beds	
2.	Hard brittle gray limestone in beds up to 12 in., few fossils	7
1.	Arkose, variously red or green .....to....	8
	Precambrian.	

Bed 2 (Tremblay) is poorly fossiliferous, but the following forms were collected:

Streptalasma corniculum (Hall), *Tetradium cellulosum* (Hall), *Rhinidictya* sp., *Ctenodonta* cf. *C. nasuta* (Hall), *Hormotoma canadensis* Ulrich and Scofield.

This faunule affords no clear picture of the age of the beds, but on the basis of the *Tetradium* they may be referred to the

Lowville. This reference is tentative only, but loosely fasciculate *Tetradium* are not usual in the higher Black River formations in Ontario and Quebec, although they recur higher in the Trenton.

Bed 3 (lower Simard) is so massive that although fossils seem to be rather common, it is very difficult to collect identifiable forms. I am sure that prolonged intensive collecting would increase this list of the forms which I found:

Streptalasma corniculum Hall, *Calapoecia* sp., *Rhinidictya* sp., *Rafinesquina* sp., *Zygospira* sp.

Ctenodonta sp., *Lophospira milleri* (Hall), *Helicotoma* sp., *Holopea* sp., *Hormotoma canadensis* Ulrich and Scofield, *Ceraurus* sp., *Bumastus* sp., *Eobronteus oratus* Sinclair, small smooth ostracods.

In bed 4, (Simard) silicification has made collecting and identification simpler, and the following list can be given:

Camarocladia sp., *Receptaculites* sp., *Stromatocerium rugosum* Hall, *Halysites quebecensis* Lambe, *Streptalasma* cf. *S. corniculum* Hall, *Lichenaria* cf. *L. typa* Winchell and Schuchert, Bryozoa spp.

Camarella sp., *Rhynchotrema* sp., *Zygospira* sp., *Helicotoma planulata* Salter, *Hormotoma gracilis* (Hall), *H. bellicincta* (Hall), H. *canadensis* Ulrich and Scofield, *Liospira vitruvia* (Billings), *Lophospira milleri* (Hall), *Omospira* sp.

Actinoceras sp., *Michelinoceras* ? sp., *Gorbyoceras* ? sp., *Westonoceras* sp., *Oncoceras* ? spp., *Sinclairoceras haha* Flower, *Eobronteus* sp., *Bumastus* sp., *Illaenus* sp., *Isotelus* sp., *Raymondites* sp., frequent ostracods.

The aspect of these two faunas is decidedly Black River, in particular recalling that of the well-known beds at Paquette Rapids on the Ottawa River.

Bed 5 (Simard) is so poorly exposed that its fauna may well be united with that of no. 6, from which it differs only in the common occurrence of *Halysites* and *Hesperorthis:*

Receptaculites occidentalis Salter, *Stromatocerium rugosum* Hall, *Favistella halli* (Nicholson), *Halysites quebecensis* Lambe, *Streptalasma* cf. *S. robustum* Billings, *Monotrypa* sp., *Prasopora* sp., *Rhinidictya* sp.

Dinorthis pectinella (Emmons), *Hesperorthis disparilis* (Conrad), *Petrocrania* sp., *Platystrophia amoena* McEwan,

Plectorthis trentonensis Foerste, *Rafinesquina alternata* (Emmons), *Resserella rogata* (Sardeson), *Rhynchotrema increbescens* (Hall), *Sowerbyella sericea* (Sowerby), *Strophomena* sp., *Zygospira recurvirostris* (Hall).

Clionychia sp., *Ctenodonta* sp., *Cyrtodonta affinis* Ulrich, *C canadensis* Billings, *C. obesa* Ulrich, *Orthodesma subnasutum* (Meek and Worthen).

Bucania sp., *Clathrospira subconica* (Hall), *Cyclonema* sp., *Ecculiomphalus trentonensis* (Billings), *Fusispira nobilis* Ulrich and Scofield, *F. subfusiformis* (Hall), *Helictomoa planulata* Salter, *Holopea parvula* Ulrich and Scofield, *H. pyrene* Billings, *H. supraplana* Ulrich and Scofield, *Hormotoma bellicincta* (Hall), *H. gracilis* (Hall), *Hyolithes* sp., *Liospira vitruvia* (Billings), *Lophospira milleri* (Hall), *L. perangulata* (Hall), *Omospira* sp., *Pterotheca* sp., *Sinuites cancellatus* (Hall), *Tetranota bidorsata* (Hall), *Trochonema umbilicatum* (Hall), *Tropidodiscus subacutus* (Ulrich).

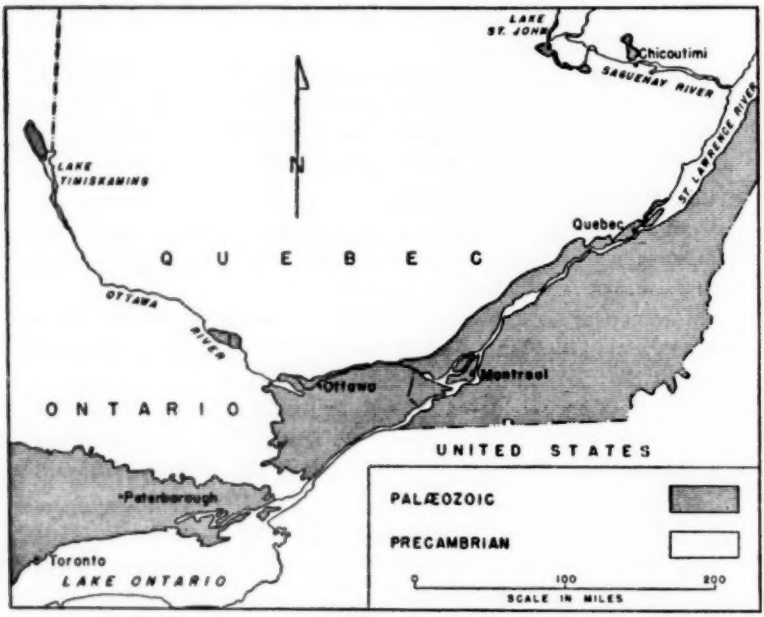

Fig. 1. Areas of outcrop of Paleozoic rocks in eastern Ontario and Quebec.

Bumastus sp., *Calliops* sp., *Flexicalymene senaria* (Conrad), *Illaenus americanus* Billings, *Isotelus* spp., *Proetus* s. l., spp., *Raymondites* cf. *R. spiniger* (Hall), *Leperditia* sp., *Technophorus* sp., *Lichenocrinus* sp.

If the beds below are doubtful in position these are not, for here we have the Paquette Rapid fauna well developed. There are some discrepancies, the presence of *Halysites*, the presence of *Fusispira* rather than *Subulites*, but these are minor, and there can be no doubt that this series of beds is at least approximately equivalent to the Ottawa Valley beds.

From the stripping, bed 7, were collected: *Streptalasma corniculum* Hall, *Prasopora* sp., *Rhinidictya* sp., *Dinorthis* sp., *Hebertella frankfortensis* Foerste, *Lingula* sp., *Öpikina* sp., *Platystrophia felis* Sinclair, *Rafinesquina alternata* (Emmons), *Resserella rogata* (Sardeson), *Rhynchotrema increbescens* (Hall), *Strophomena billingsi* Winchell and Schuchert, *S. thalia* Billings, *Zygospira recurvirostris* (Hall).

Cyrtodonta obesa Ulrich, *Bucania* sp., *Helicotoma planulata* Salter, *Holopea appressa* Ulrich and Scofield, *Hormotoma canadensis* Ulrich and Scofield, *H. gracilis* (Hall), *Liospira americana* (Billings), *Lophospira milleri* (Hall), *L. perangulata* (Hall), *Tetranota bidorsata* (Hall), *Trochonema umbilicatum* (Hall), *Flexicalyene senaria* (Conrad), *Lichenocrinus* sp.

This fauna is entirely noncommittal. The aspect is as much Trenton as Black River, but it does not recall any particular part of the Trenton. The fossils one would expect in the Rockland or Hull are missing. The exposure is too small to justify any definite assignment.

Chute aux Galets.—The beds exposed along the Shipshaw River show the following continuous section:

3.	Black shale, resting on a somewhat roughened surface (Gloucester shale)	25 ft.
2.	Rather thick-bedded limestones, weathering rubbly (Galet beds)	15
1.	Thin-bedded dense limestones with thick (up to 3 in.) soft shaly partings (Shipshaw beds)	60

The black shales were not examined in detail, except near their contact with the limestones. Their thickness is taken from McGerrigle's estimate. Besides the fossils which he lists from these beds, my small collection includes *Triarthrus spinosus*

and *Schizocrania filosa*, species which make a Gloucester correlation of the beds quite certain. Indeed the lithology and fauna are such that specimens from Chute aux Galets and Ottawa cannot be distinguished.

The upper 15 feet of the limestone is separable from the underlying beds, although no sharp line can be drawn between them. There is a gradual diminution in the amount of shale as the section is ascended, and the limestone, which below is platy, becomes dirtier and rubbly. More important, there is a faunal difference. The upper beds are not abundantly fossiliferous, but they include common *Cyclospira* (of which only one specimen was found in the lower beds), and *Illaenus americanus* as a common fossil. The fauna of these Galet beds is:

Streptalasma corniculum Hall, *Conularia trentonensis* Hall, *Prasopora* sp., *Rhinidictya* sp., *Cyclospira bisulcata* (Emmons), *Platystrophia amoena* McEwan, *Rafinesquina deltoidea* (Conrad), *Resserella rogata* (Sardeson), *Sowerbyella sericea* (Sowerby), *Strophomena* sp.

Bucania sp., *Hormotoma trentonensis* Ulrich and Scofield, *Liospira americana* (Billings), *Flexicalymene senaria* (Conrad), *Illaenus americanus* Billings, *Isotelus* sp.

The lower shaly Shipshaw beds (1) are abundantly fossiliferous, and yielded:

Ischadites iowensis (Owen), *Pasceolus globosus* Billings, *Receptaculites occidentalis* Salter, *Diplograptus* sp., *Streptalasma corniculum* Hall, *Conularia trentonensis* Hall, *Cornulites flexuosus* (Hall), *Serpulites* sp., *Prasopora* sp., *Rhinidictya* sp., *Subretepora* sp.

Craniops trentonensis (Hall), *Cyclospira bisulcata* (Emmons), *Dinorthis pectinella* (Emmons), *Hesperorthis tricenaria* (Conrad), *Öpikina* sp., *Parastrophina hemiplicata* (Hall), *Plaesiomys subquadrata* (Hall), *Platystrophia amoena* McEwan, *Plectorthis trentonensis* Foerste, *Pseudolingula major* Ruedemann, *Rafinesquina alternata* (Emmons), *R. camerata* (Conrad), *R. deltoidea* (Conrad), *R. praecursor* Raymond, *Resserella rogata* (Sardeson), *Rhynchotrema increbescens* (Hall), *Sowerbyella sericea* (Sowerby), *Strophomena billingsi* Winchell and Schuchert, *S. thalia* Billings, *S. trilobita* Owen, *Vellamo trentonensis* (Raymond).

Ctenodonta alta (Hall), *Endodesma gesneri* (Billings), *Whitella* sp., *Bucania* sp., *Clathrospira subconica* (Hall),

Fusispira nobilis Ulrich and Scofield, ***F. subfusiformis*** (Hall), *Holopea symmetrica* Hall, ***Hormotoma salteri*** Ulrich and Scofield, ***H. trentonensis*** Ulrich and Scofield, ***Liospira americana*** (Billings), ***Lophospira oweni*** Ulrich and Scofield, ***Omospira*** sp., ***Phragmolithes compressus*** Conrad, ***Sinuites cancellatus*** (Hall), ***Subulites subelongatus*** d'Orbigny, ***Tetranota bidorsata*** (Hall), ***Trochonema umbilicatum*** (Hall).

Bumastus sp., ***Ceraurinus*** cf. ***C. confluens*** Barton, ***Ceraurus dentatus*** Raymond and Barton, *C.* sp., ***Dimeropyge lucifer*** Sinclair, ***Encrinurus*** sp., ***Flexicalymene*** sp., ***Hemiarges*** sp., ***Illaenus americanus*** Billings, ***Isotelus*** sp., ***Proetus*** s.l., sp., ***Leperditia*** sp., ***Cheirocrinus*** sp., ***Glyptocystites batheri*** Sinclair, ***G. regnelli*** Sinclair, ***Hudsonaster*** sp., ***Lichenocrinus*** sp.

The fauna of these beds is of strikingly Sherman Fall aspect, although such elements as the ***Fusispira*** and ***Rafinesquina camerata*** would be surprising in these beds in Ontario (but not in Quebec, where the Rosemount member of the Montreal formation bears many ***Fusispira***). I consider the Shipshaw as of late Sherman Fall age, being succeeded above by definitely Cobourg (Galet) beds. It may be that the whole thickness is Cobourg, but I do not think so. Until the fauna can be studied critically there is no point in pretending to greater precision.

It will be noted that I find no beds at Chute aux Galets which overlap the section at Tremblay. Rather, there seems to be a long gap in the record, with beds of Hull and lower Sherman Fall age unrepresented. There is room for them in the covered area between Tremblay and Chute aux Galets, but also a possibility that they were not deposited in this area at all.

OUTLIERS AT LAKE ST. JOHN

The distribution of the Ordovician beds along the south side of Lake St. John has been described by Dresser. I was not able to establish a continuous section in this area, mainly because the level of the lake had been raised some 30 feet by power developments, and the old shore sections drowned. It is regrettable that Dresser gave no section, since apparently when he studied the district the shore could be traversed. I was able to see at various places small exposures showing (1) the contact of the limestones with the underlying Precambrian, (2) the contact with the overlying black shales, and (3) a number of

small quarries in beds of middle Trenton aspect. My notes, therefore, form three disconnected parts.

Precambrian contact.—Small exposures along the railroad north of Roberval were described and illustrated by Dresser (1916, pl. 4B). The basal sediments are rather thin-bedded limestones, with no residual weathered material between them and the irregular Precambrian surface. Hollows in the surface were filled rather rapidly, and at a height of a foot or so above the contact the bedding of the limestones is essentially level. Fossils are few, but collecting conditions are not good, and this poverty may be illusory. Some beds have poorly silicified brachiopods. I did not see here any beds which could have yielded the slabs with finely silicified mollusks which were collected by Richardson and by Dresser and are preserved in the Geological Survey collections in Ottawa. Presumably they came from a shore exposure which is not now accessible. At the railroad cut I found in the 7 feet of beds exposed there:

Receptaculites occidentalis Salter, *Streptalasma corniculum* Hall, *Rhinidictya* sp., *Dinorthis* cf. *D. alternata* Wilson, *D.* sp., *Rafinesquina* sp. aff. *R. deltoidea* (Conrad), *Rhynchotrema* sp.,

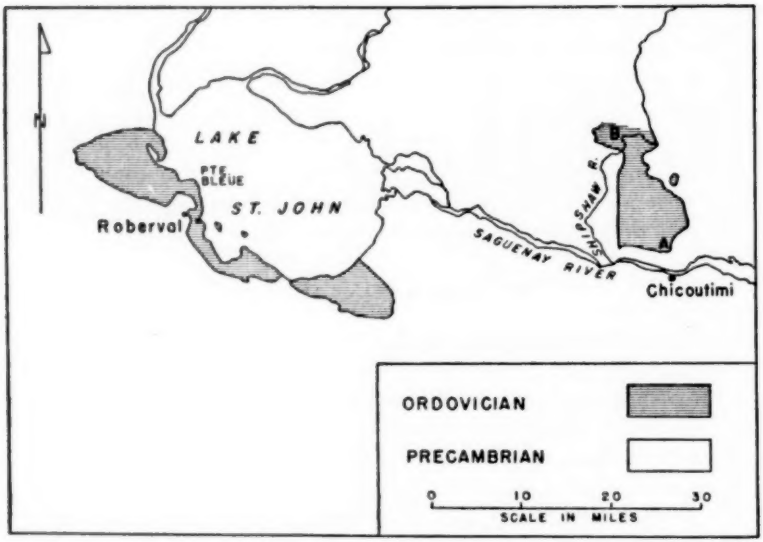

Fig. 2. Lake St. John, Quebec, and vicinity, showing the distribution of the outcrops discussed in the paper. North of Chicoutimi the Tremblay quarry area is marked A, Chute aux Galets is marked B.

Zygospira recurvirostris (Hall), *Bucania* sp., *Trochonema canadensis* Ulrich and Scofield.

No *Halysites* was seen. This small faunule does not permit a definite assignment, but I feel that it belongs above the *Halysites* beds at Tremblay.

Middle Trenton.—A series of small quarries about two miles north of Roberval (lots 3-5 of range A) expose about 47 feet of beds which seem to be Middle Trenton. The same beds are repeated several times by small faults, but I did not see them in contact with other beds at any place. The lithology is a rather argillaceous limestone but with the clayey material so disseminated that on fresh surfaces the appearance is of a very thick-bedded limestone, in beds up to 5 feet thick. Only on very long weathering do these beds break down to a softer rubbly mass. The fauna of the three quarries studied is essentially the same, and is combined here:

Hindia parva Ulrich, *Ischadites iowensis* (Owen), *Pasceolus globosus* Billings, *Receptaculites occidentalis* Salter, *Climacograptus* sp., *Diplograptus* sp., *Aulopora trentonensis* Winchell and Schuchert, *Protaraea vetusta* (Hall), *Streptalasma corniculum* Hall.

Conularia trentonensis Hall, *Cornulites flexuosus* (Hall), *Metaconularia* sp., *Corynotrypa* sp., *Helopora* sp., *Monotrypa* sp., *Prasopora* sp., *Rhinidictya* sp., *Subretepora* sp.

Craniops trentonensis (Hall), *Cyclospira bisulcata* (Emmons), *Glyptorthis* sp., *Hesperorthis disparilis* (Conrad), *H. tricenaria* (Conrad), *Öpikina* sp., *Parastrophina hemiplicata* (Hall), *Petrocrania* sp., *Plaesiomys meedsi* (Winchell and Schuchert), *P. ulrichi* (Foerste), *Platystrophia amoena* McEwan, *Pseudolingula major* Ruedemann, *Rafinesquina alternata* (Emmons), *R. camerata* (Conrad), *R. deltoidea* (Conrad), *R. normalis* Wilson, *R. praecursor* Raymond, *R. robusta* Wilson, *R. salmoni* Wilson, *Resserella rogata* (Sardeson), *Rhynchotrema increbescens* (Hall), *R. intermedia* Wilson, *Sowerbyella sericea* (Sowerby), *Strophomena filitexta* Hall, *S. scofieldi* Winchell and Schuchert, *Trematis ottawaensis* Billings, *Vellamo trentonensis* (Raymond), *Zygospira recurvirostris* (Hall).

Ctenodonta intermedia (Ulrich), *Cyrtodonta obesa* Ulrich, *Orthodesma gesneri* (Billings), *Whitella* sp., *Archinacella trentonensis* Billings, *Bellerophon clausus* Ulrich, *Bucania* sp., *Clathrospira subconica* (Hall), *Cyclonema* sp., *Ecculiomphalus tren-*

tonensis (Billings), *Fusispira angusta* Ulrich and Scofield, *F. nobilis* Ulrich and Scofield, *Gyronema liratum* Ulrich and Scofield, *Hormotoma bellicincta* (Hall), *H. gracilis* (Hall), *H. trentonensis* Ulrich and Scofield, *Liospira americana* (Billings), *L. angustata* Ulrich and Scofield, *L. micula* (Hall), *Lophospira milleri* (Hall), *L. perangulata* (Hall), *Phragmolithes compressus* Conrad, *Proplina unguiformis* Ulrich and Scofield, *Sinuites cancellatus* (Hall), *Subulites nanus* Ulrich and Scofield, *S. subelongatus* d'Orbigny, *Trochonema umbilicatum* (Hall).

Achatella sp., *Acrolichas* aff. *A. trentonensis* (Conrad), *Bumastus* sp., *Calliops* sp., *Calyptaulax* sp., *Ceraurinus* cf. *C. confluens* Barton, *Ceraurus dentatus* Raymond and Barton, *C. matranseris* Sinclair, *Dimeropyge lucifer* Sinclair, *Encrinurus* spp., *Eomonorachos* sp., *Flexicalymene senaria* (Conrad), *Hemiarges* sp., *Homotelus* sp., *Illaenus americanus* Billings, *Isotelus* sp., *Otarion* sp., *Proetus* s. l., spp., *Remopleurides* aff. *R. striatulus* Walcott, *R.* aff. *R. linguatus* Ruedemann, *Sphaerocoryphe* sp.

Leperditia sp., *Eurychilina* sp., *Primitia* sp., *Tetradella* sp., *Glyptocrinus* sp., *Glyptocystites* sp., *Hudsonaster narrawayi* (Hudson), *Lepidocoleus* sp., *Lichenocrinus* sp., *Promopalaeaster* sp., *Pleurocystites* sp.

Taken as a whole, this is a typical later Sherman Fall fauna, but the details are puzzling. In spite of the abundant representation of the trilobites, the harpeids and the leonaspids are completely missing, and those groups which have been studied in detail seem to differ specifically from the Trenton forms of New York and adjacent areas. Again, some elements point to Cobourg affinities. Until more critical determinations can be made, nothing more definite can be said than that the fauna seems to be "Middle" Trenton. Certainly, it is the same as the fauna of the Shipshaw beds at Chute aux Galets.

Black shale contact.—Several small exposures along the lake shore exhibit the contact of the black shales on the limestones. The best section seen was:

3. Black shale with *Triarthrus glaber, T. spinosus*. Very typical Gloucester shale.
2. Thick-bedded limestone, with few fossils other than large straight cephalopods. The upper surface, at the contact with the shale, strikingly uneven ..... 15 ft.

1. Somewhat thinner-bedded limestone, with rubbly or thin shaly interbeds 18

Although well exposed, the beds seemed sparingly fossiliferous, only the following species being found in bed 1:

Ischadites cf. *I. iowensis* (Owen), *Streptalasma corniculum* Hall, *Climacograptus* sp., *Dinorthis* cf. *D. subquadratus* (Hall), *Glyptorthis* sp., *Rafinesquina* cf. *R. camerata* (Conrad), *Strophomena* sp., *Sowerbyella* cf. *S. semiovalis* Wilson, *Parastrophina* sp., *Hormotoma trentonensis* Ulrich and Scofield, *Isotelus* cf. *I. gigas* DeKay.

This is a Cobourg type of assemblage, so far as it goes, and taken in conjunction with its position, I think we may look on these beds as Cobourg, and equivalent to the Galet beds at Chute aux Galets.

In summary, we have at Lake St. John beds which are apparently equivalent to the Shipshaw and the Galet, but no certain representation of the lower beds seen at Tremblay. It must be remembered that *Halysites quebecensis* was originally described from Lake St. John, and so (unless its range be greater than suspected) beds equivalent to the Simard must occur somewhere along the lake, although I did not see them.

REGIONAL IMPLICATIONS

Having established a tentative section for the Saguenay outliers, with representation of beds from Lowville to Gloucester age, some implications must be mentioned although they cannot be developed here.

(1) The succession described here is on the whole that found at Ottawa and in central Ontario near Peterborough. Although details differ, the resemblance of the Galet to the Cobourg, of the Shipshaw to the beds in Ontario identified as Sherman Fall, is striking. Now, this Sherman Fall-Cobourg sequence is not found at Montreal, nor in the St. Lawrence Valley below Montreal. Although the Montreal Tetrauville formation can be correlated with the Cobourg with some assurance, still it differs lithologically and faunally from the Ontario beds and the resemblances are very general ones. The succeeding Terrebonne is unknown in Ontario, unless indeed it be equivalent to the Collingwood, which it does not resemble except in position. As one moves northeast in the St. Lawrence both these thick limestone formations are replaced by shales, just as they are

replaced by shales as one moves south into the Champlain-Hudson valley. If the late Trenton beds at Peterborough, Ottawa and Lake St. John are similar lithologically and faunally, as they are, where was the connection? Not to the south, for we know the Trenton there, and it is different. I think we must imagine a continuous belt of carbonate facies running roughly parallel to the St. Lawrence, but 100 miles or so to the northwest of it. Judging by the distribution of clastics, this was an offshore belt, with the source of detrital materials south and east of the St. Lawrence. Obviously, adjustments are needed in our thinking about Laurentia in Trenton time. I understand that a number of other areas of Trenton limestones are known still further within the Shield, although their descriptions have not been published. I do not think we are nearly ready to make maps of the Ordovician seas in this part of the continent.

(2) It will be noted that at both Lake St. John and Chute aux Galets the contact of the Gloucester on beds of Cobourg age is seen, and there is no trace of the Collingwood. The whole question of the distribution and relationships of the Collingwood is a large one, and cannot be considered here. The present study only eliminates one possible direction of encroachment of the Collingwood fauna.

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GLACIAL FEATURES OF COOK COUNTY, MINNESOTA*

ROBERT P. SHARP

ABSTRACT. Reliable evidence has been found for only two ice invasions of Cook County, although the glacial history of this area must have been much more complex. The Rainy lobe moved into the area from the north and covered the entire county during the Cary substage of the Wisconsin. It was succeeded by the Superior lobe which moved west-southwest along the Superior trough during the Mankato substage of the Wisconsin and covered a narrow belt along the southern margin of Cook County. These advances were not contemporaneous because Superior lobe red clayey till rests upon Rainy lobe brown sandy till, and red lake clays derived from the Superior lobe were deposited over areas earlier occupied by the Rainy lobe.

Lakes associated with these ice sheets have occupied parts of Cook County at the following times: (1) prior to or during early stages of the Rainy lobe advance, (2) during or shortly after recession of the Rainy lobe, (3) prior to or contemporaneous with advance of the Superior lobe, (4) during the climax and recessional phases of the Superior lobe, and (5) after all ice had disappeared from Cook County but still lay in neighboring regions.

Contrary to earlier reports, end moraines are not present in Cook County. The most obvious glacial deposits are principally irregular accumulations of sand and gravel formed in association with masses of stagnant ice isolated in low areas during recession of the Rainy lobe. Sixteen eskers, the largest 20 miles long, have been identified and mapped. The internal constitution and structure and the external form, orientation, and arrangement suggest that they were formed largely by streams flowing chiefly in ice tunnels but possibly to some extent in ice-walled gorges. The esker streams flowed directly on the subglacial floor in some places, and in others they were not more than 10 to 20 feet above the glacier's base. Wide gaps in esker ridges are thought to represent places where the streams flowed across still thicker ice, so that the deposits eventually became superglacial and were dispersed. The internal "anticlinal" structure of at least one esker is considered to be a primary depositional feature, but the exact mode of origin of this structure is not understood. Crude drumloid ridges were formed in southwestern Cook County by the Rainy lobe.

INTRODUCTION

General statement.—Cook County lay directly in the path of continental ice masses which repeatedly moved south out of Canada and west-southwest down the Superior trough during the Pleistocene. Its glacial history must have been long and complicated, but the complete record is not preserved.

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Evidence has been found for only two ice invasions, both of Wisconsin age.

The Rainy lobe (Elftman, 1898, p. 108) moved down from the north and covered all of Cook County. The exact age of this advance is not known, but it is unquestionably Wisconsin, most likely about middle Wisconsin (Wisconsin III or Cary). The Superior lobe (Leverett, 1929, p. 7) came from the east-northeast, followed the Superior trough, and covered only a narrow strip of land along the southern edge of Cook County. Its age is presumed to be late Wisconsin (Wisconsin IV or Mankato).

Judging from Leverett's map (1932, pl. 1), he attributes all the drift mapped in Cook County to the Superior lobe. This is clearly refuted by evidence cited in following pages which shows that the Mankato Superior lobe extended at most 5.5 miles north of the present Lake Superior shore. Furthermore, this evidence does not support the contemporaneous occupation of parts of Cook County by the Rainy and Superior lobes as formerly suggested by Elftman and Leverett, nor are the features in Cook County previously mapped as end moraines believed to be such.

This study constitutes part of a geological investigation of Cook County made largely by F. F. Grout, and sponsored by

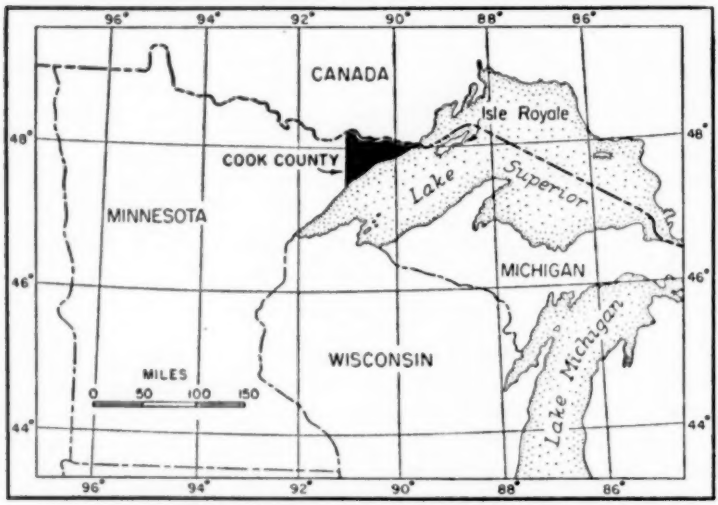


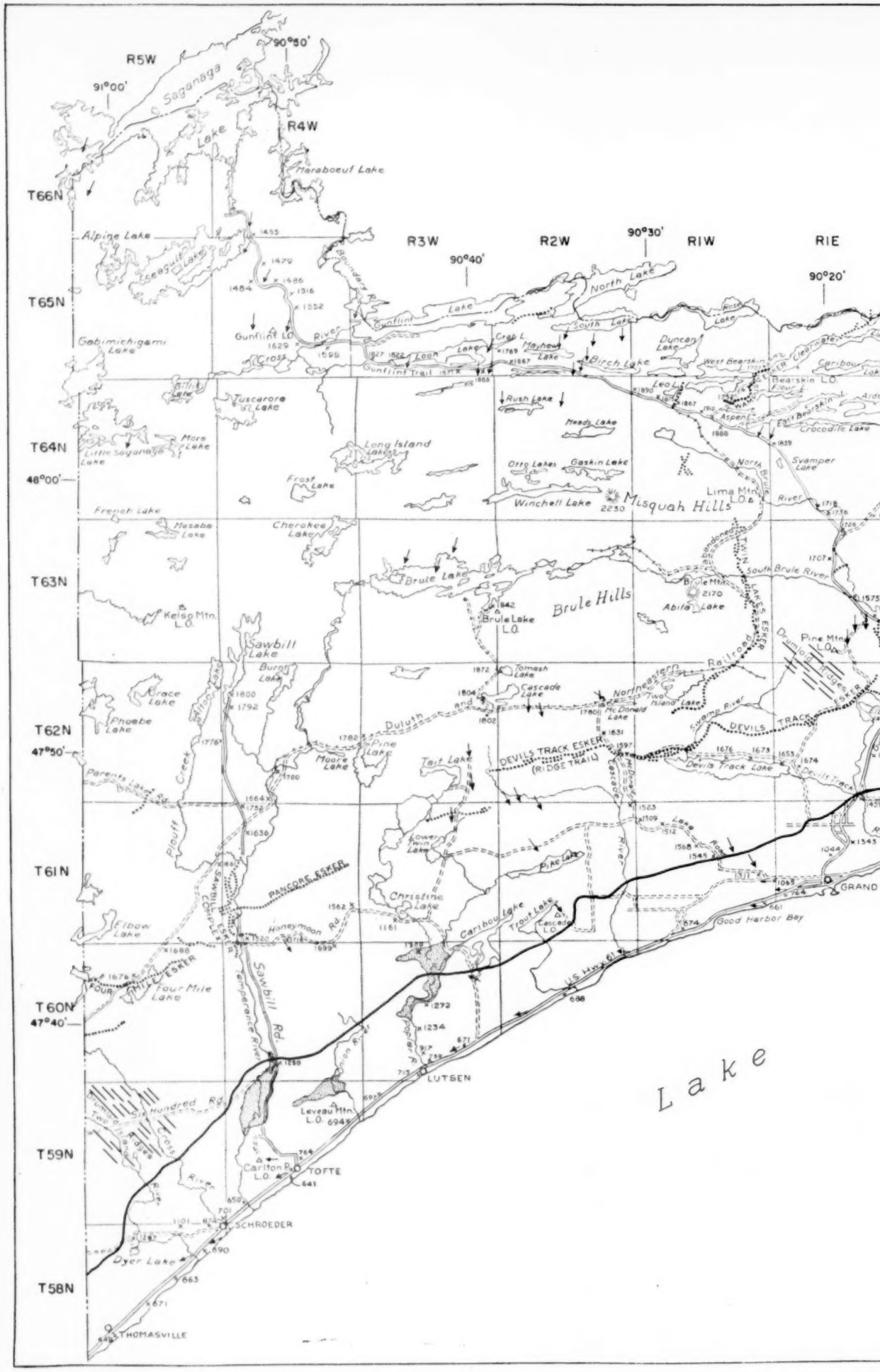
Fig. 1. Location map of Cook County.

the Minnesota Geological Survey. Appreciation is expressed for the aid, advice, and suggestions of Grout, G. M. Schwartz, and H. E. Wright. The Graduate School of the University of Minnesota provided funds for purchase of air photographs. Field work on the Pleistocene features of Cook County covered a period of approximately 3½ months in the summers of 1946 and 1947. Considering the size of the area and the vegetative cover, this must be classified as a reconnaissance study.

Physical setting.—Cook is the northeasternmost county (fig. 1) of Minnesota (47° 55′ N., 90° 30′ W.). Its triangular shape is determined by the eastward convergence of the Canadian border and the north shore of Lake Superior. About 17 percent of the total area of 1680 square miles in the county is said to be water-covered (Leverett and Sardeson, 1917, p. 50), but recent air photos suggest that this is an underestimate. The area is hilly, with local relief up to several hundred feet and altitudes ranging from 602 feet, the surface of Lake Superior, to 2230 feet in the Misquah Hills, the highest point in Minnesota. Were the water drained from Lake Superior, this figure would be increased nearly 1000 feet, and it is the total bedrock relief that is of principal significance in the glacial relations.

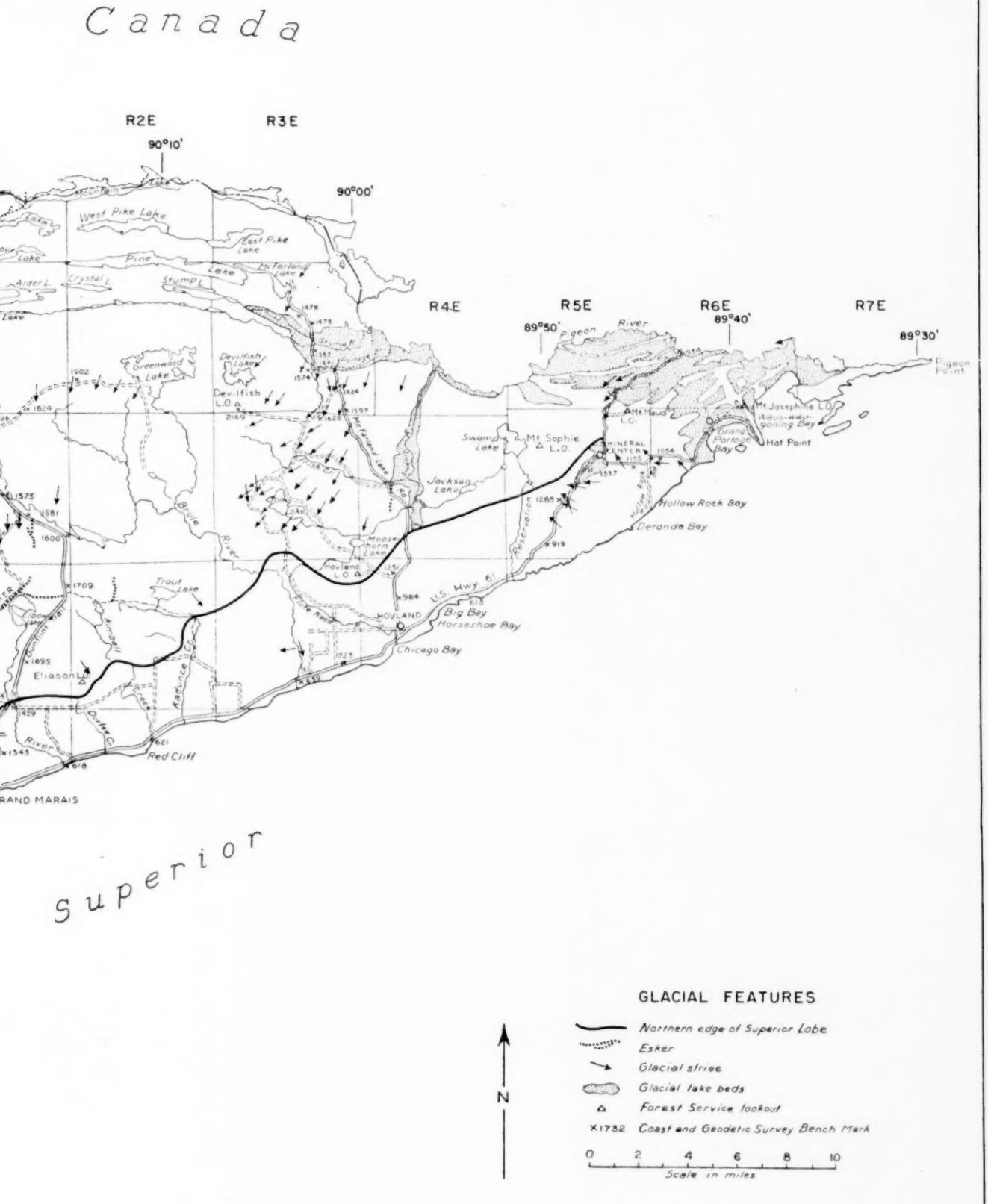
Physiographically, the county is divisible into three units: (a) the Coastal Hills along the shore of Lake Superior rising to elevations as much as 1200 feet above lake level within a few miles inland; (b) the Interior Upland, a rolling surface of slight relief lying north of the Coastal Hills at elevations between 1400 and 1900 feet with occasional ridges and groups of hills exceeding 2000 feet; and (c) the Northern Ridges and Valleys, or the Rove area, which features long lake-filled valleys at altitudes of 1500 to 1800 feet lying beside cliff-faced cuesta ridges in some instances almost 500 feet higher.

The bedrock of Cook County is wholly Precambrian, with Saganaga granite, Ely greenstone or Knife Lake series, Pokegama quartzite and Gunflint iron formation in the northwest; Rove formation slate, graywacke, and diabase in the north and northeast; largely Duluth gabbro with redrock differentiates and Keweenawan volcanics in the Interior Upland; and chiefly Keweenawan volcanics and red sedimentary



NOTE: SVAMPER LAKE (T. 64 N., R. 1 E.) SHOULD READ SWAMPER LAKE.

Fig. 2. Map of glacia



lacial features in Cook County.

beds, with diabase intrusives, in the Coastal Hills (Thiel, 1947, figs. 15-16).

The number of lakes in Cook County, excluding swamps, probably exceeds 1000. The deepest, aside from Lake Superior, is Saganaga Lake at 240 feet, and the largest lying wholly within the county is Brule Lake, which covers 4062 acres.[1] Several of these lakes have multiple outlets.

This is essentially a sparsely settled "wilderness" area with only a few roads and trails. It is covered by a dense second-growth forest of deciduous and coniferous trees with an undergrowth of other plants (Butters and Abbe). The available maps are small scale and planemetric only, but air-photo coverage is now available for the entire county.

GLACIAL EROSION

The principal effects of glacial erosion in Cook County are small markings on ice-scoured outcrops, including polish, striations, grooves, chattermarks, friction cracks and small joint-controlled excavations; and larger landscape features such as rounded outcrops, glacially scoured and plucked hills and ridges, and basins and valleys excavated in the bedrock. Only those erosional features useful in determining direction of ice movement are considered here.

Most Rainy lobe striations bear within 30° on either side of south (fig. 2). The various trends appear to reflect the influence of local topographic situations and were undoubtedly controlled to some degree by the direction of slope of the glacier's surface according to principles best presented by Nye (1952, p. 92). Striations show that flow within the Rainy lobe changed from a southerly to a southeasterly direction in the southern part of Cook County. This was probably due to the southeasterly slope of the land approaching the Lake Superior trough. Superior lobe striations show movement west-southwest down the Lake Superior trough, with a slight tendency to fan out northwestward onto the adjacent upland. This fanning probably expresses the control on direction of flow exerted by a relatively steep outward slope of the ice surface along the margin of the lobe.

[1] Data on lakes kindly supplied by Professor Samuel Eddy of the University of Minnesota. An excellent publication on Minnesota lakes is also now available (Zumberge, 1952).

Of the several types of friction cracks described by Harris (1943) only crescentic fractures were found in this region, an especially fine display being on an outcrop of Northern Light gneiss in the northeast arm of Saganaga Lake just north of Cook County. Many glaciated rock surfaces have joint-controlled scars, a few inches to one or two feet across and an inch or two deep, from which a mass of rock has been plucked. Most scars are defined by one joint plane dipping gently with or against the direction of flow and by one or more steep joints transverse to the direction of flow (fig. 3). Modification of these scars by glacier scour shows that they were formed beneath moving ice, or at least were overridden by it, and the asymmetry of scouring indicates the direction of ice movement.

Most of the prominent bedrock hills and ridges in Cook County show some degree of asymmetry related to glacial plucking and oversteepening on their lee slopes. Carlton Peak near Tofte has steep sides to both the south and west,

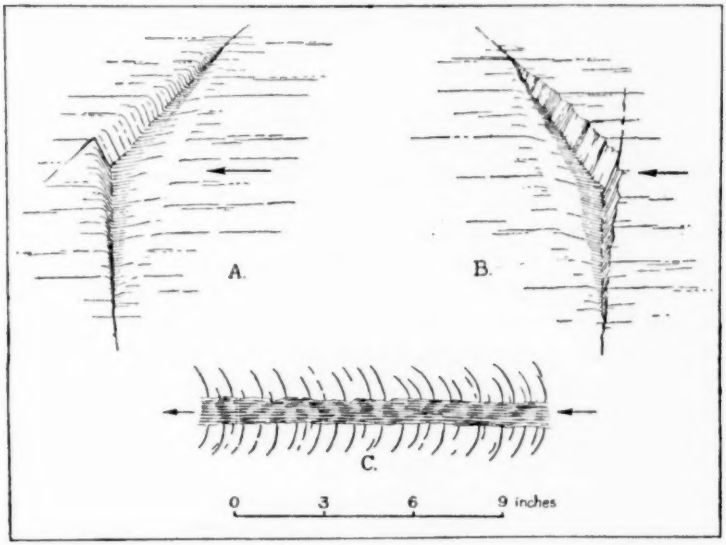


Fig. 3. Field sketches of minor features on glaciated bedrock surfaces. A—joint-controlled excavation with ice moving against gently dipping face. B—same with ice moving against gently dipping face. C—vertical view of glacial gouge and associated crescentic fractures. Arrows indicate direction of ice movement.

perhaps reflecting initial plucking by the south-moving Rainy lobe and later plucking by the west-flowing Superior lobe. Exceptions to the usual asymmetry are afforded by cuesta-like strike ridges underlain by massive diabase sills in the Rove area of northern Cook County (Grout and Schwartz, 1933; Zumberge, 1952, p. 24-30). Here rock structure is the predominant influence, and the steeper slopes of these ridges are to the north, facing against rather than with the direction of ice movement. This came about because glacier plucking and sapping of fissile sedimentary beds underlying the south-dipping diabase sills created or at least maintained a steep face to the north.

GLACIAL DEPOSITION

Brown sandy till (Rainy lobe, Cary or Wisconsin III).—The most extensive glacial deposit in Cook County is a discontinuous mantle of brown sandy till left by a south-moving ice mass which covered the entire county. Good exposures of this till can be seen along Sawbill road (T. 60 N., R. 4 W.), along Gunflint Trail south of Swamper Lake (T. 64 N., R. 1 E.), and along the abandoned right-of-way of the Duluth and Northeastern Railroad east and west of Sawbill road (Ts. 61 and 62 N., Rs. 4 and 5 W.).

Brown is the prevailing color with minor and local variations to grayish, blackish, and reddish hues. The till is stony with a sandy matrix. Less than 5 percent has a matrix of clay or silt, probably obtained by local incorporation of lake deposits. Stones are composed of gabbro, intermediate and red-rock differentiates of gabbro, diabase, taconite, jasper, iron formation, greenstone, granite, porphyritic granite, syenite, gneiss, schist, amygdaloid, felsite, rhyolite, basalt, basalt porphyry, quartzite, graywacke, red and white arkosic sandstones, and quartz-pebble conglomerate. The largest boulders are 15 feet in diameter. The till shows many compositional variations reflecting the nature of underlying or nearby bedrock. Near exposures of the Rove formation, for example along the road just south of McDonald Lake, it consists largely of fragments of argillite with some quartzite, graywacke, and diabase. On or near belts of redrock it is composed of redrock stones set in a sandy matrix of ground up redrock. Till in which 90 percent of the stones are redrock has been observed to grade within

one mile to till in which 90 percent of the stones are Keweenawan volcanics.

The till is loose and uncemented in most exposures, displays considerable oxidation, and shows some decomposition and disintegration of rocks especially susceptible to weathering, chiefly the olivine gabbros and diabases. The soil mantle is thin, and its profile immature. Average thickness of the till probably does not exceed 15 feet, and in many places it is even thinner although wells suggest greater thicknesses in favorable spots (Thiel, 1947, p. 98-102). Along the old railroad grade northeast of Sawbill road are exposures of brown

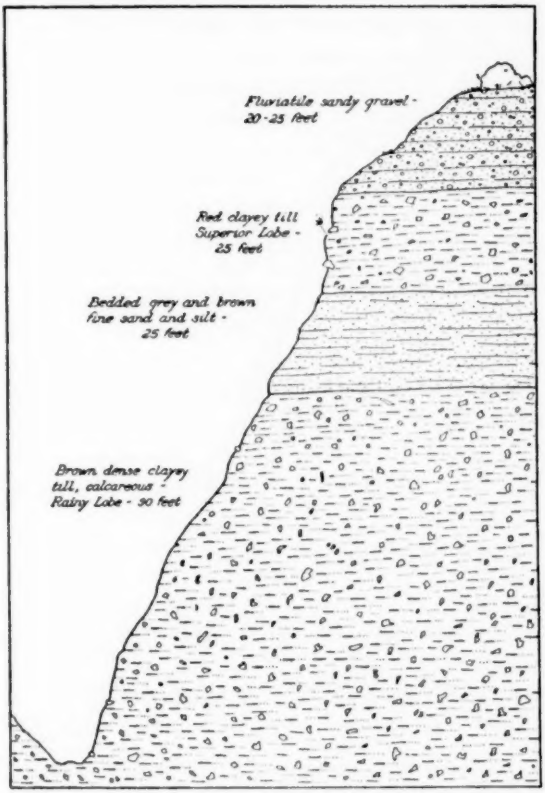

Fluviatile sandy gravel - 20-25 feet
Red clayey till Superior Lobe - 25 feet
Bedded grey and brown fine sand and silt - 25 feet
Brown dense clayey till, calcareous - Rainy Lobe - 30 feet

Fig. 4. Field sketch of drift section exposed in east wall of Brule River gorge just below Pothole Falls (south center of Sec. 22, T. 62 N., R. 3 E.).

sandy till 35 to 50 feet thick, and near Pothole Falls on Brule River (Sec. 27, T. 62 N., R. 3 E.) 93 feet of brown calcareous clayey till are exposed beneath sandy silts and red clayey till (fig. 4). This brown clayey till is provisionally included with the brown sandy till, although it might be older.

Leverett (1932, pls. 1 and 3) appears to attribute the brown sandy till to the Superior lobe, although Elftman (1898, p. 92-96, 108-109) had earlier attributed it to the Rainy lobe. Elftman's interpretation is considered correct for the following reasons: (1) Striations (fig. 2) associated with the brown sandy till trend mostly within 30° of north and south. That the movement was southward is shown by the asymmetry of glaciated bedrock hills, by minor joint-controlled depressions, by friction cracks, and most definitely by composition of the till. (2) Numerous exposures along the south edge of the county display brown sandy till underlying red clayey till.[2] This relation is somewhat obscured by red illuvial clay which has been carried down into the brown till from the overlying deposits, but it has been confirmed repeatedly, and in places glacifluvial beds (figs. 4 and 5) lie between the two tills. It is clear that the brown and red tills are two distinct deposits and that the brown till was formed by a south-moving ice mass, for which Elftman's (1898, p. 108) term, the Rainy lobe, is accepted. This advance probably occurred in the Cary (Wisconsin III) substage, but treatment of age relations is taken up following consideration of the red clayey till.

Both Elftman (1898, pl. 11) and Leverett (1929, p. 28-29; 1932, pl. 1) mapped end moraines in the area covered by the Rainy lobe. Careful restudy of the features so mapped shows that they are either bedrock ridges thinly mantled with till or accumulations of glacifluvial material developed in association with isolated bodies of stagnant ice. No definite end moraines were found in all of Cook County.

Red clayey till (Superior lobe, Mankato or Wisconsin IV).—The Superior lobe moved west-southwest along the Lake

[2] This stratigraphic relation appears to have been discovered by Franklin B. Hanley whose work in Cook County was cut short by death in 1944. Hanley's field notes could not be located, but the caption on one of his lantern slides at the University of Minnesota identified red till resting on brown till in Cook County.

Superior trough and left a discontinuous mantle of red clayey till over a belt 3 to 5.5 miles wide on the southern margin of Cook County (fig. 2). This till is best preserved in the broad valleys of the Coastal Hills, and good exposures can be seen along Caribou Lake road (Sec. 11, T. 60 N., R. 3 W.), along U. S. Highway 61 at Poplar Cemetery 1.5 miles west of Grand Marais, along a road east of Poplar River (center of Sec. 21, T. 60 N., R. 3 W.), and on Gunflint Trail north of Grand Marais (Secs. 9 and 10, T. 61 N., R. 1 E.). Most of this till is red or pink, but in places near the base it is brownish pink and locally brown owing to incorporation of brown lake clay or brown sandy till. Younger red lake clay has been laid down over much of the area covered

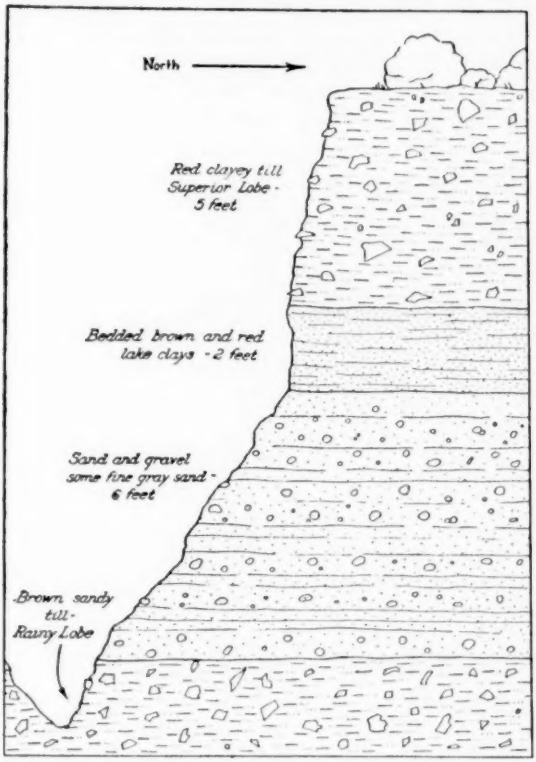


Fig. 5. Field sketch of drift section exposed on Kimball Creek (south edge Sec. 28, T. 62 N., R. 2 E.).

by the Superior lobe, and every exposure of red clay-rich material raises the problem of distinguishing between clayey till and lake clay. This is especially difficult where the lake clay has been affected by some mass movement.

In its typical form the till is compact and contains 60 to 80 percent red clay, both as matrix and as discrete fragments. Sandy and stone-rich facies are rare. Most are 0.5 to 4 inches in diameter, but some attain 2 feet. The most abundant rock types are rhyolite, basalt, redrock, diabase, greenstone, granite, gneiss, white and red arkosic sandstone, conglomerate, and a little cream-colored limestone. The only stones lithologically distinct from those in the brown sandy till are the limestone and a dense uniform greenstone, both of unknown source. Fragments of Keweenawan sandstone, conglomerate, basalt, and rhyolite are considerably more abundant than in the brown sandy till.

At Big Bay and on Poplar River the red clayey till is calcareous. On the crest of a well-drained ridge along Poplar River it is leached of carbonates to a depth of 2.5 feet. In general, the visual weathered zone on the till is only 16 to 18 inches thick, consisting of 6 to 12 inches of dark humic soil underlain by brownish oxidized material. The modest depth of alteration may reflect both the impervious nature of the till and a relatively short interval of weathering.

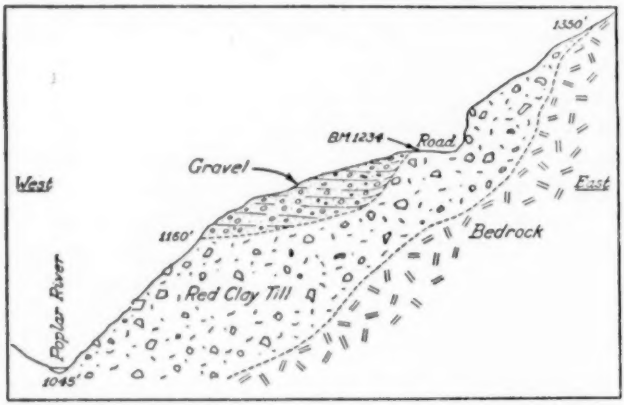


Fig. 6. Interpretative sketch of relations on Poplar River suggesting a thick till filling in a preglacial rock gorge (south center of Sec. 21, T. 60 N., R. 3 W.).

PLATE 1

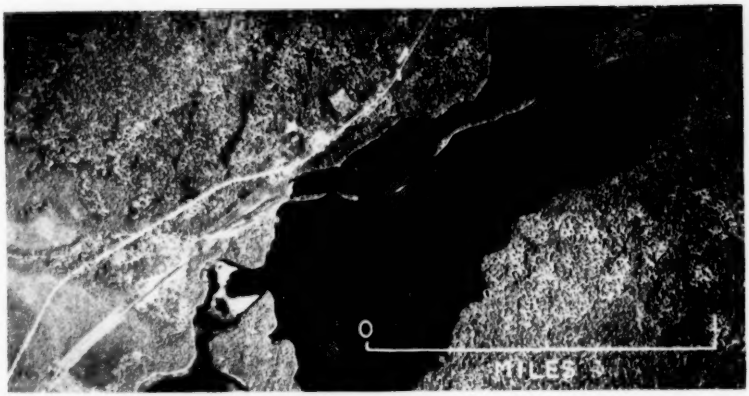


Fig. 1. Fourmile esker in Fourmile Lake (T. 60 N., R. 5 W.). Top is north. U. S. Forest Service air photo.

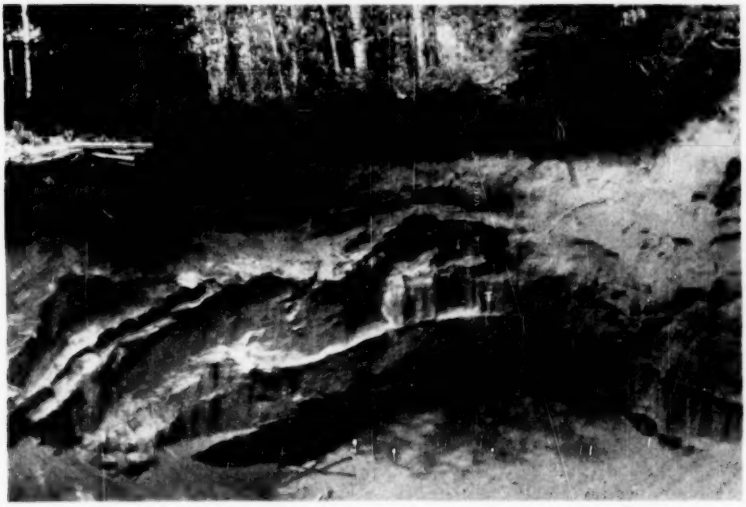


Fig. 2. "Anticlinal" bedding in Wampus esker near West Bearskin Lake (NW. ¼ Sec. 1, T. 64 N., R. 1 W.). Material is largely coarse sand and fine pebble gravel.

Fig. 1. Pancore esker between Pancore and Missouri lakes; Poplar River in middle (T. 61 N., R. 4 W.). Top is north. U. S. Forest Service air photo.

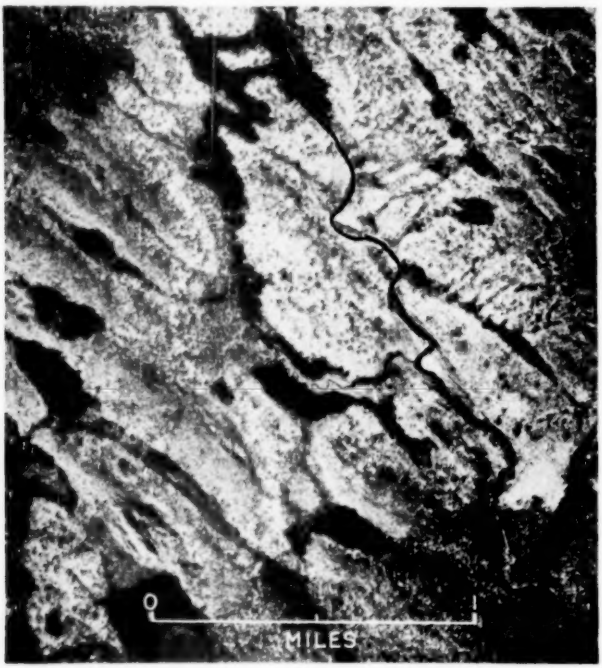


Fig. 2. Drumloid topography along Cross River in T. 59 N., R. 5 W. Top is north. U. S. Forest Service air photo.

In most exposures, the red clayey till is 5 to 15 feet thick, but locally as much as 25 feet is seen, and somewhat greater thicknesses are recorded in some wells (Thiel, 1947, p. 98-102). Along Poplar River (Sec. 21, T. 60 N., R. 3 W.) the exposures and relations depicted in figure 6 suggest that a preglacial rock gorge was filled to a depth of 200 to 250 feet by red clayey till.

The lack of sharp topographic features on the Superior lobe till is probably due to its high content of clay which gives great mobility when wet and facilitates spreading into low inconspicuous forms. The topography, composition, and thickness of the till deposits northwest of Grand Marais (Sec. 15, T. 61 N., R. 1 W.) are more suggestive of an end moraine than any other accumulation in Cook County, but even so they do not constitute a convincing example of such a feature.

Age of glaciations.—The abundant lakes, the freshness of the glacial forms, especially those composed of glacifluvial materials, and the lack of deep or advanced weathering preclude an age greater than Wisconsin and favor mid-Wisconsin or younger for the two drifts in Cook County. The distribution of Cary (Wisconsin III) drift farther south indicates that Cook County could hardly have escaped being glaciated during this substage, and similar relations suggest that it could have been partly or completely overrun by Mankato ice. H. E. Wright (1952) has collected evidence farther west and south in Minnesota indicating Cary and Mankato advances of both the Superior and Rainy lobes. Ruhe (1952, p. 400, fig. 2) shows all of Cook County as covered by Mankato ice, but this appears to be an unsupported generalization.

Even though the red clayey till overlies the brown sandy till, both might be either Cary or Mankato. However, the preferred interpretation assigns the brown sandy till to the Cary and the red clayey till to the Mankato. This interpretation hinges on the interval of time between deposition of these two tills, a point on which the local evidence is indicative but not compelling. It is as follows:

No perceptible evidence of weathering has been found on the brown till where overlain by red till, but considerable incorporation of brown material in the basal red till suggests glacial erosion along this contact; and Thwaites (1943, p. 136) could find little evidence of weathering between Valders

(Mankato) and Cary till in Wisconsin. Weathering at this contact would be decisive, but under these circumstances its absence is not fatal to the preferred interpretation. More positive evidence is furnished by red lake clays, derived from the Superior lobe, which were deposited over an area in northern Cook County formerly occupied not only by the Rainy lobe but also by a lake receiving debris from that lobe as it shrank. The Rainy lobe must have receded north of the Canadian border before the Superior lobe was in a position to shed meltwater and debris into this area, the basins of the Pigeon and Stump rivers.

These relations suggest that the culminating phases of the two ice lobes were not closely contemporaneous. The Cary-Mankato interval appears to have been relatively brief in Iowa, southern Minnesota, and the Dakotas (Flint, 1950, p. 1461; Ruhe, 1952), and it was probably even shorter in areas closer to the center of glaciation, such as Cook County, because of the shorter time between the waning and waxing phases of successive glaciations. H. E. Wright (personal communication) and associates have obtained evidence farther west in Minnesota suggesting that ice which deposited the brown sandy till in Cook County was distinctly earlier than other ice masses considered to be Mankato on independent evidence. For these reasons a Cary age is favored for the brown sandy till in Cook County. Leverett (1929, p. 21, 38-39) has assigned a Mankato date to the red clayey till in Cook County largely on the basis of regional studies, and nothing incompatible with this dating has been discovered during the present work. If these age assignments are correct, evidence for a Superior lobe of Cary age (H. E. Wright, 1952) has not been found here, and behavior of the Rainy lobe in relation to Cook County during Mankato time remains unknown.

Glacial lake deposits.—Parts of Cook County have been occupied at various times by extramarginal glacial lakes which received meltwater and debris from the ice. Their former existence is demonstrated primarily by beds of red clay, locally calcareous and in some places resting upon older beds of brown non-calcareous clay and silt. Such lake deposits are exposed principally in the valleys of the Stump, Swamp, and Pigeon rivers; the Mineral Center region; the lower valleys of the

Poplar, Onion, and Temperance rivers; and in areas immediately bordering Lake Superior (fig. 2).

In the broad flat valley of Stump River 1 to 2 feet of red clay rest with sharp contact on brown clay layers interbedded with silt in a manner faintly suggestive of varves. Total thickness of the brown beds is not known, but it is at least several feet. Around the margins of this valley the brown beds rise to 1415 feet altitude, but the red clay attains only 1385 feet. Red clay beds extending up to 1385 feet altitude on Swamp River and its tributaries were deposited in an arm of the same lake that flooded Stump River. Exposures and excavations on Swamp River were too shallow to show whether these red beds are also underlain by brown lake deposits.

The water body occupying the valleys of the Stump and Swamp rivers was named Lake Omimi by Elftman (1898, p. 104), but Leverett (1929, pl. 2) considered it to be part of Lake Duluth. The interpretation favored here is that these valleys were flooded at two different times, the brown beds being deposited in a body of water nourished by the Rainy lobe, namely Lake Omimi, and the red beds originating in a later lake dammed and fed by the Superior lobe. This second lake was not part of Lake Duluth, however, for that water body did not rise to a level high enough to flood Stump River (Sharp, 1953, p. 116). The nature of the dam impounding Lake Omimi is not known, but it may have been a detached and stagnant mass of Rainy lobe ice somewhere along the Pigeon River. It seems unlikely that the Superior lobe formed the dam because the generally sharp contact between the red and brown beds suggests that red material was not being contributed to the lake while the brown beds were being deposited.

Much of the lower Pigeon River-Grand Portage region is covered by lake deposits. Not all parts of this region shown on the map (fig. 2) as lake beds have been visited in the field, but their elevation, appearance on air photographs, and relation to areas known to be underlain by lake beds favor this interpretation. About 200 yards south of Pigeon River on U. S. Highway 61, at 980 feet altitude, 20 feet of red clay overlie with sharp contact 30 feet of brown silty clay. Southward along U. S. 61 the red clay decreases in thickness and rises to 1400 feet. As on Stump River, the brown silty clays are attributed to an earlier lake nourished by the Rainy lobe

and the red clay to a later lake receiving debris from the Superior lobe. The considerable thickness of red beds at the Pigeon River locality is probably due to additional deposition in a succession of lakes marginal to the Superior lobe, including Lake Duluth and stages of Lake Algonquin.

Mineral Center (T. 63 N., R. 5 E.) is situated in a broad basin covered with at least 7 feet of red lake clay. This clay is massive to poorly bedded, and contains calcareous nodules as well as occasional stones and patches of sand. Since the upper limit of these clay beds, 1350 feet, is above any known or postulated level of Lake Duluth in this vicinity, they are assigned to a somewhat older lake held at a higher level along the margin of the Superior lobe. The water body at Mineral Center appears to have been independent of the lakes on Stump and Pigeon rivers.

In the southwestern part of the county, lake beds are exposed in the valleys of Poplar, Onion, and Temperance rivers on the northern side of the Coastal Hills (fig. 2). For the most part these accumulations are thin and consist entirely of red clay, but at Oxbow Camp on Temperance River (Sec. 33, T. 60 N., R. 4 W.) the 50 feet of red clay exposed contains interbeds of fine brownish sand and silt near the top. The upper limit of lake beds is 1309 feet on Poplar River, 1250 feet on Onion River, and 1240 feet on Temperance River. These elevations are above known or postulated shorelines of Lake Duluth at these places (Sharp, 1953, p. 116), so the deposits are attributed to marginal lakes formed when the valley mouths were dammed by the Superior lobe. Meltwater carried red debris north into the lakes which must have attained their maximum extent after some recession of the ice had occurred, for the beds lie largely in areas formerly covered by the Superior lobe. The brown sand and silt near the top of the Temperance River section were probably reworked from Rainy lobe drift.

The land immediately bordering Lake Superior was submerged by Lake Duluth and its successors, and within this area are scattered deposits of red lake clay up to 10 feet thick.

The high content of red lake clay in Superior lobe till indicates that the ice moved over areas mantled by lake beds older than any red beds discussed so far, and on Kimball

Creek is an exposure (fig. 5) showing 2 feet of red clay beneath red clayey till. Such clay beds may have been deposited in a proglacial lake related to a receding ice mass such as Martin's (1932, p. 453-457) Lake Rouge or Lake Sioux or Murray's (1953, p. 153) Lake Keweenaw which formerly occupied the Superior basin. However, there is no proof that the beds on Kimball Creek were not deposited in a smaller lake bordering the advancing Superior lobe. Clay-rich phases of the Rainy lobe till also suggest incorporation of earlier lake beds, and brown silty clays are exposed beneath the brown till along Gunflint Trail just south of Devils Track River.

From the above, it appears that late Pleistocene glacial lakes occupied parts of Cook County at the following times: (1) prior to or during the early stages of the Rainy lobe advance, (2) during or shortly after recession of the Rainy lobe, (3) prior to or contemporaneously with advance of the Superior lobe, (4) during the climax and recessional phases of the Superior lobe, and (5) after all ice had disappeared from Cook County but still lay in neighboring regions (Lake Duluth et al.).

Irregular glacifluvial accumulations formed in association with stagnant ice.—Deposits of glacifluvial material with extremely irregular topography cover parts of central and northern Cook County. The most accessible are along Sawbill road (T. 61 N., R. 4 W.), at the west end of Devils Track Lake (T. 62 N., R. 1 W.), in the vicinity of Elbow Lake (T. 62 N., R. 1 E.), and in some of the northern valleys, particularly near Leo, Hungry Jack, Aspen, Flour, West Bearskin, and Clearwater lakes (Ts. 64 and 65 N., Rs. 1 W. and 1 E.). These deposits stand out in the landscape and on air photos because of their topographic peculiarities, and in the past some of them were mapped as end moraines (Elftman, 1898, pl. 11; Leverett, 1932, pl. 1). As shown in the following discussion this interpretation is considered erroneous, and similar accumulations in other areas likewise have caused confusion (J. W. Goldthwait, 1938; Mannerfelt, 1938).

In grain size the glacifluvial materials range from sand to boulders, and the stones are the same as in the brown sandy till. Bedding, sorting, and rounding of particles range from poor to good, and in most places the deposits are clearly waterlaid. Locally a layer of bouldery unbedded detritus containing

erratics up to 15 feet in diameter mantles the glacifluvial deposits. This mantle is generally thicker and coarser than a slopewash or a lag concentrate, and it is probably composed of superglacial detritus dumped onto the glacifluvial beds from nearby higher masses of ice.

Topographically the glacifluvial materials form a complex of swells and swales, knobs and kettles, and short ridges without consistent trend, pattern, or character. The ridges bifurcate, intersect, broaden or thin, and change direction abruptly. Some of the knobs are strung together as beaded ridges. The local relief is 20 to 50 feet and slopes are steep, mostly 20 to 30 degrees. Lowland areas are the favored topographic setting for glacifluvial accumulations.

The origin of these deposits is pictured as follows. As the waning Rainy lobe thinned by down-melting (Ahlmann, 1938, p. 337; Flint, 1942) masses of ice lying in lowland areas became isolated from the main body owing to the relatively strong relief of the subglacial floor. These isolated masses stagnated and wasted away in situ. A heavy cover of superglacial debris developed and was moved about on the extremely irregular surface that seems to characterize most masses of stagnant ice. Both ponded and running water were abundant, and drainage from nearby slopes and outwash from other ice masses probably brought considerable debris to these areas. The volume of material seems too large to have come solely from the stagnant ice.

Eventually, the ice melted away completely in a few places, and here the deposits were laid down in their final resting place upon the ground. Such spots were probably originally small, irregular in outline, largely unconnected, and partly or wholly enclosed by walls of ice. As wastage continued they became larger, even more irregular, and were integrated to some degree with each other. Eventually, as the ice grew thinner and the deposits thicker, or at least more concentrated, large masses of ice were completely buried in debris. Subsequent melting of these buried masses caused collapse and slumping of the overlying and bordering deposits. The knobs, swells, and ridges of the present topography therefore represent areas where the ice disappeared first and the thickest deposits accumulated, and the swales and kettles represent areas where ice lingered longer and the deposits were thinner. The ridges

were formed in elongated depressions which may have been controlled to some degree by structures in the ice.

The suggestion has been made that some glacifluvial deposits resembling those in Cook County are of subglacial origin (Ver Wiebe, 1927, p. 177; Andersen, 1931, p. 613; Priehäusser, 1938, p. 109). According to this concept deposition occurs under thin stagnant ice that is essentially afloat in its own meltwater, and the topography developed is supposedly a cast of the bottom of the overlying ice mass. The large volume of the Cook County deposits, their sharp topographic forms with slopes at the angle of repose, and the lack of internal structures conformable to the supposedly molded forms suggest that this mode of origin does not apply here. Thwaites (1943, p. 106) was also unsuccessful in finding indications of a subglacial origin for similar deposits in Wisconsin.

Eskers.—The sixteen eskers mapped in Cook County are grouped largely in two belts, one extending west-southwesterly to the western edge of the county, and a second extending southward from the Canadian border and almost joining the first (fig. 2). Without much imagination these two belts and most of the component eskers within them could be joined into a single stream-like system. Notable eskers are Fourmile (pl. 1, fig. 1), Pancore (pl. 2, fig. 1), Devils Track (fig. 7), Twin

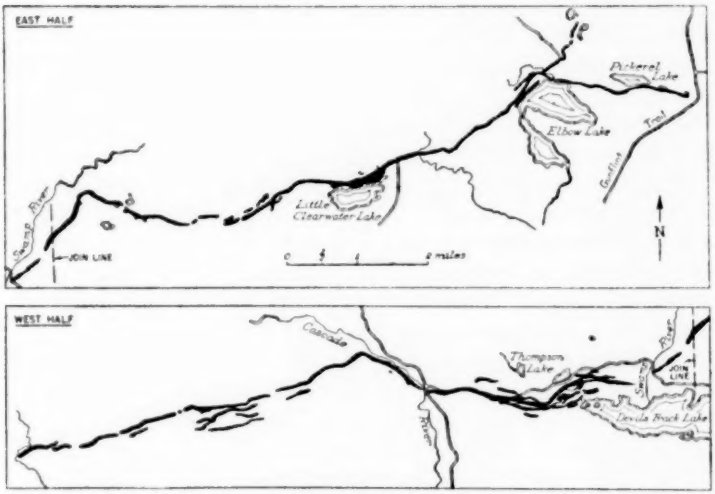


Fig. 7. Map of Devils Track esker.

Lake (fig. 8) and Wampus (fig. 9). No Cook County eskers are situated on or cross topographic eminences, and none was found in areas covered by the Superior lobe. Those of the west-southwest belt trend almost at right angles to the direction of movement within the Rainy lobe, as indicated by nearby striae, and thus provide an exception to the generalization that eskers are usually parallel to the direction of glacier flow (Flint, 1947, p. 151).

Gravel is the major constituent of Cook County eskers, but sand is abundant in the matrix and locally composes individual beds. In places Wampus esker consists largely of fine to coarse sand with a little pebble gravel (pl. 1, fig. 2). Boulders up to 4 feet in diameter are embedded in the esker deposits, and larger boulders, some 15 feet across, are scattered about on the surface. A structureless mantle of bouldery detritus, 2 to 3 feet thick, on the crests and slopes of many Cook County eskers looks like ablation debris deposited from overlying or nearby ice. Similar mantles in other areas have been interpreted as a slopewash or lag concentrate (Brown, 1931, p. 476).

Stones in the esker gravel are subangular and their wide range in composition indicates that they have been derived in large part if not wholly from Rainy lobe ice or the brown sandy till. Eskers are usually composed of reworked ice-carried debris (Trefethen and Trefethen, 1944), and there is no indication here that much of the material was derived by direct erosion of the bedrock by the esker-forming stream (Hershey, 1897, p. 244). Sorting and bedding are poor except in a few layers of pebbly gravel and of sand. Small-scale cross bedding was observed, but larger deltaic foresets are lacking. The pseudo "anticlinal" structure characteristic of many eskers is shown here in a few instances. Mostly it appears to be related to surficial slumping or creep, but Wampus esker shows an internal "anticlinal" structure which is not the product of surficial processes (pl. 1, fig. 2).

Most Cook County eskers are 20 to 30 feet high, some attain 50 feet, and a few are 80 to 90 feet high. Lengths of 0.5 to 3 miles are usual, but Twin Lake esker is at least 10 miles long, and Devils Track esker can be traced with confidence for 20 miles. Side slopes range from 20° to 35°, depending upon material, and asymmetry in cross section is not uncom-

mon. The steeper slope is not consistently on the same side in neighboring eskers or in different places on the same esker. The internal bedding of Wampus esker in one locality dips 34° west and 25° east, roughly conformable with its external form. Air photos and field observations suggest that the steeper

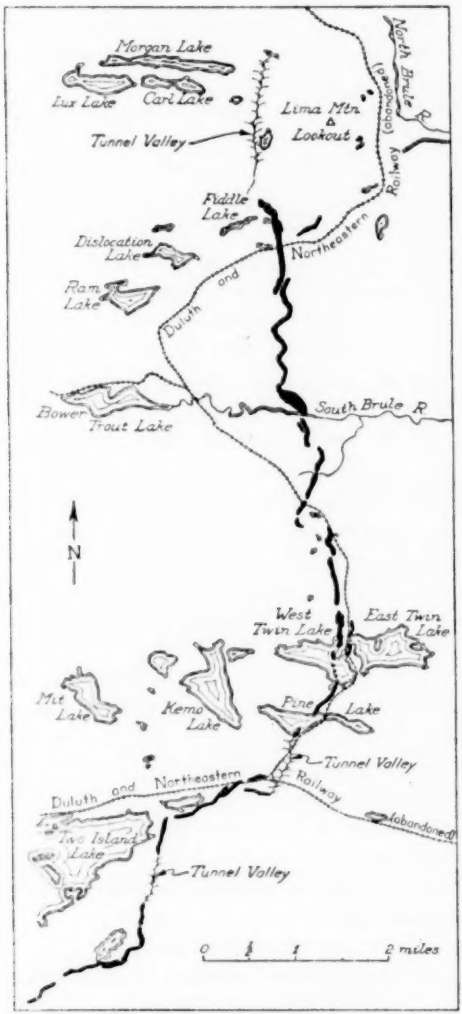


Fig. 8. Map of Twin Lake esker.

side marks a tendency toward lateral shifting by the esker-forming stream, for it is located largely on the outsides of curves.

Most of these eskers are irregular in plan. Neither the nearly straight 2-mile stretch of Pancore esker (pl. 2, fig. 1) nor the symmetrical meandering curves near the north end of Twin Lake esker (fig. 8) are typical. Some single ridges bifurcate and reunite thus inclosing a deep, steep-sided depression, a habit well shown by some Canadian eskers (Lang, Bostock, Fortier, 1947, A 4218-32). Single ridges also locally give way to a braided or reticulate network of ridges or to irregular glacifluvial deposits with knob and kettle topography. At curves some eskers display several subparallel ridges as though the esker stream underwent a succession of lateral shifts. This is not universal nor are subparallel ridges limited to curves. Stream-like tributary branches also exist, for example, on Devils Track esker at Elbow Lake (fig. 7) which, unless it is a barbed junction, indicates that the esker stream flowed westward. The Sawbill assemblage, consisting of an irregular arrangement of short subparallel esker ridges of north-south and east-west trend, is best termed an esker complex.

Stream-cut channels interrupting the course of an esker were found only along Twin Lake esker (fig. 8), which displays narrow gorges cut 10 to 30 feet into low bedrock prominences in three places along its course. These appear to have been formed by the same stream that built the esker, and in some areas they are known as esker valleys or tunnel valleys (Andersen, 1931, p. 616). Devils Track is the only Cook County esker with an extensive marginal fosse or esker trough.

The origin of eskers has been long debated, and many aspects of the problem remain unsettled. It is clear that the various features to which the name esker has been applied were not all formed in exactly the same way. Deposition by streams of glacial meltwater in intimate association with the ice is accepted as the principal mechanism. However, the details of the process and the various conditions under which it operates are not yet agreed upon.

According to one concept, deposition occurs at the mouth of a stream debouching into ponded water at the margin of the ice. As the ice front recedes a succession of elongated delta segments is built and joined together to form an esker (DeGeer,

1910, p. 243-245; W. B. Wright, 1914, p. 41-43; Flint, 1947, p. 153). In Scandinavia associated varves are said to show that each delta segment is an annual deposit, and experiments (Hanson, 1943, p. 451-452) suggest that esker-like forms can be built in this manner. Accumulation at the margin of a receding ice front without benefit of ponded water has also been advocated (Hershey, 1897, p. 244; Trowbridge, 1914, p. 216).

In the United States a generally favored concept has been one of deposition from heavily laden streams flowing in tunnels beneath stagnant or relatively inactive ice. Deposition is variously attributed to overloading, to choking of the mouth of the stream by proglacial deposition (Russell, 1893, p. 240-242), to an opposed gradient on the glacier's floor (Chadwick, 1928, p. 926), and to ponding of water beneath the ice (Andersen, 1931, p. 613-614; Flint, 1930, p. 628-629; Lewis, 1949, p. 318; Meier, 1951, p. 293). Minor compressional deformation of the esker deposits suggests that the ice forming the tunnel need not have been entirely inactive (Flint, 1942, p. 130-131).

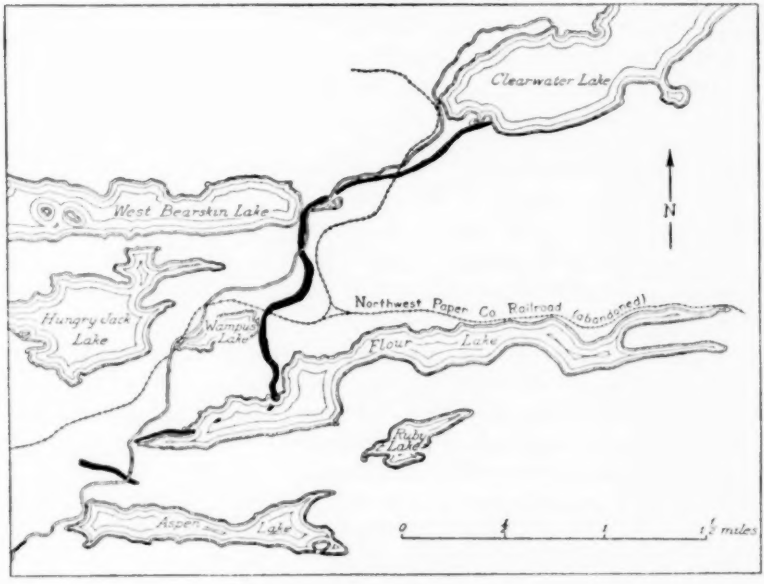


Fig. 9. Map of Wampus esker.

The formation of eskers by englacial (Alden, 1925, p. 54) or superglacial streams (Winchell, 1884, p. 668-669; Crosby, 1902) has not been widely accepted, although Hartshorn (1952) feels that some eskers associated with the Malaspina Glacier may have had a superglacial origin. The difficulty here lies in transferring superglacial or englacial accumulations to the ground without seriously altering or destroying them, a problem to be considered in more detail shortly. Wilson (1939, p. 126-127) proposes that certain Canadian eskers were formed by streams, initially superglacial, which cut their channels through the thin edge of an ice sheet and reworked the debris in and at the bottom of the ice. This type of esker is formed in an ice-walled gorge open to the sky and is supposedly built headward as the ice front recedes. Open ice-walled gorges are also favored by others (Dryer, 1901, p. 128; L. Goldthwait, 1939, p. 114-115), and often the evidence does not permit one to distinguish between deposits originating in subglacial tunnels and open ice-walled gorges (Thwaites, 1943, p. 110).

Eskers associated with existing glaciers throw some light on the problem of origin, and Washburn's (1941, p. 222) fine photograph of an esker extending outward from the Woodworth Glacier, Alaska, furnishes the following information although confirmation by study on the ground is needed: (1) The ice front has receded in normal fashion without evidence of stagnation in the immediate vicinity of the esker, although isolated bodies of stagnant ice have been left in nearby areas. (2) Erosional channels cut in the ground moraine between the end of the esker and the present ice edge appear to have been formed by the same stream that built the esker but not necessarily at the same time. (3) Braiding and bifurcation of the esker occurs at curves and in places where its course makes a large angle with the direction of ice flow, suggesting that the channel was displaced by stream erosion or by glacier movement. (4) Deposition of the esker appears to have been related to the building of an ice-contact outwash fan at an earlier stabilized position of the glacier front and to have occurred at the same time. (5) Progressive deposition of the esker along a receding ice front seems unlikely, in view of the lack of any outwash whatsoever on the surrounding ground moraine.

It seems clear that deposition of this esker must have taken place in an ice tunnel or ice-walled gorge at or almost at the bottom of the glacier. A subglacial tunnel is favored to some degree by the lack of any ice-walled gorge of comparable extent in the present glacier. Eskers associated with existing glaciers in the Wind River Mountains (Meier, 1951, p. 293) and in Norway (Lewis, 1949, p. 318) are also attributed to deposition in subglacial tunnels. Origin of Cook County eskers is now considered in the light of the above discussion.

The features of all Cook County eskers are similar enough so that great differences in the mode of origin seem unlikely. The possibility that the eskers of the west-southwest belt were formed as kame-like ice-contact deposits strung laterally along the margin of an ice sheet is not supported by any known evidence except their orientation, and it is opposed by lack of any external form or internal structure consistent with such an origin. Accumulation in successive segments along an ice front receding east-northeasterly also seems unlikely since independent evidence for an ice front of proper orientation receding in that direction is lacking. Although the west-southwest belt of eskers does not occupy a conspicuous lowland, it does lie north of and parallel to the higher Coastal Hills, suggesting that topographic control of subglacial drainage determined its orientation.

Accumulation of Cook County eskers in ice-walled gorges or in ice tunnels is suggested by: (1) the stream-like patterns with tributary branches; (2) the bifurcated, reticulated, and braided arrangements; (3) the multiple subparallel ridges at curves; (4) the orientation of the transverse asymmetry; (5) the localization in lowland areas; (6) the rock gorges along the course of Twin Lake esker; (7) the local heavy mantle of coarse ablation debris; and (8) the steep reverse faults in one esker suggesting mild compression by slightly active ice. However, the "up-hill down-dale" course of some eskers, frequently cited as proof of accumulation in subglacial tunnels, is not found in Cook County.

If the above features are accepted as indicating deposition in ice gorges or tunnels, then one must ask whether the streams flowed wholly on the subglacial floor, or wholly or in part upon ice somewhat above the bottom of the glacier. This introduces the problem of transferring a superglacial or en-

glacial accumulation to its final resting place on the ground without seriously modifying or entirely destroying its external form and internal structure. In the instance of deposits formed close to the bottom of a thick glacier such a transfer may be possible because these deposits can be let down upon the ground by basal melting before surface ablation brings them into the superglacial environment. Basal melting of a temperate glacier (Ahlmann, 1948, p. 66) occurs at a rate of about ¼ inch per year. Surface ablation of bare ice may easily be as much as 10 to 20 feet a year, but a thick debris mantle can reduce this to a few inches or even less. Thus, an englacial deposit a few feet above the base of a glacier, several hundred feet thick, may have a chance of becoming subglacial before the surface of the glacier is lowered enough to make it superglacial.

However, suppose it becomes superglacial, what are the possibilities of transferring this or any other superglacial deposit in a relatively undisturbed state to a resting place upon the ground? If the mantle of superglacial debris is of such homogeneity in character and thickness that the ice surface is lowered uniformly by ablation, and if basal melting is also uniform, superglacial accumulations may arrive on the ground in a relatively undisturbed condition. Most studies of stagnant glaciers have emphasized the great differences in surface ablation from place to place, the chaotic topography, and the constant shifting of the superglacial debris. However, on the stagnant margin of the Malaspina Glacier and upon other stagnant Alaskan glaciers, a superglacial mantle of sand and gravel has attained stability sufficient for spruce trees to have taken root and to have attained an age of at least 100 years. Melting of the ice beneath this mantle must be so uniform that no tilting of the trees or serious damage to their roots results. The rate of surface melting here may be of the order of an inch, more or less, per year, and the underlying ice in most places is at least 50 to 100 feet thick. However, even these relatively stable tree-covered superglacial accumulations are not likely to reach the ground in an unaltered state, as they are currently being destroyed by the back-melting of ice faces and cliffs moving outward from the margins of ponds and stream channels cut in the ice. These observations suggest that it is possible for surface ablation to be relatively uniform, but if the underlying ice is more than perhaps 20

feet thick, it would seem that the long period of time, literally centuries, required to melt the ice allows considerable opportunity for other processes to come into action and to destroy the original form of the superglacial accumulations. Preservation of superglacial eskers underlain by no more than 10 to 20 feet of ice appears to be within the realm of possibility under favorable conditions.

Some features of Cook County eskers suggest that at least in places the streams forming them flowed over ice. One such feature is the irregular longitudinal profile of most eskers. They have marked peaks and saddles and are more undulatory than normally expected of undisturbed stream beds. Such undulations might conceivably be formed in a subglacial tunnel if the deposits were molded against an irregular ceiling, but it is easier to picture them as the product of deposition on an irregular floor of thin ice which subsequently melted away allowing some differential settling of the deposits.

More significant in this connection are the wide gaps in some esker ridges. Some gaps are clearly the result of post-depositional erosion, but many are wide, flat-floored, and show no evidence of erosion either by later transverse streams or by the esker stream itself. The suggestion is offered that these gaps represent reaches where the esker stream was underlain by ice so thick that surface ablation eventually brought the deposits into a superglacial position where they were subsequently destroyed.

The origin of the internal "anticlinal" structure displayed by some eskers is a baffling problem. In those instances where the structure is clearly surficial, it is probably the product of the surface slump and creep characteristic of most ice-contact surfaces. Where it extends deeply into the esker something more than surficial modification is indicated. In some instances the structure is probably primary rather than secondary, but no satisfactory mechanism has yet been proposed for depositing these beds with an initial anticlinal form. Perhaps this may occur in a subglacial tunnel partly filled with debris and being slowly enlarged by melting of the walls, or, if eskers are molded against the base of thin stagnant ice essentially afloat, as suggested by Andersen (1931, p. 612-616), the anticlinal structure may represent successive episodes of molded deposition as the overlying ice was floated progressively

to a higher and higher position. It is also possible that large-scale settling of a deposit filling a wide U- or V-shaped channel cut in ice is involved. This last mechanism is attractive because of simplicity, but the considerable elongation required in the uppermost beds suggests that small normal faults or thinning of the beds should be apparent. The best example of anticlinal bedding in a Cook County esker (pl. 1, fig. 2) shows no thinning of the beds and displays only steep faults, some of which have reverse displacements. This suggests that the structure is a primary depositional feature, but the exact mechanism involved in its origin remains a mystery.

In summary, Cook County eskers are thought to have been formed by streams flowing in ice-walled gorges or ice tunnels, mostly the latter, located directly on the subglacial floor in some places and no more than 10 to 20 feet above the glacier's base in others. Gaps in eskers are thought to indicate spots where the underlying ice was still thicker, so that the deposits subsequently became superglacial and were destroyed. The ice surrounding or enclosing the eskers was relatively inactive but not all of it was entirely stagnant.

Drumloid ridges.—In westernmost Cook County along the Six Hundred Road (T. 59 N., R. 5 W.) is a series of short parallel ridges trending about S. 45° E. (pl. 2, fig. 2). They are 10 to 25 feet high, 0.5 to 0.75 mile long, up to 0.25 mile wide, round-topped, equally blunt at both ends, and separated by wide swales. Some of the minor ridges lie on the flanks of larger ridges. These features occupy a broad lowland along Cross River where brown sandy till is the only material seen, but exposures are shallow. An area northwest of Elbow Lake (Sec. 5, T. 62 N., R. 1 E.) has similar but less conspicuous forms bearing roughly S. 50° E.

The first thought was that the ridges must be bedrock thinly mantled with till and that their orientation reflects a structural control. However, no exposures of bedrock have been found in the area, and the ridges are discordant to any known structural trend in this part of the county. Subsequent discovery of glacial striations bearing S. 35° E. about 8 miles to the northeast of the Cross River area suggests that the ridges are parallel to the direction of flow in the Rainy lobe. The drumloid ridges in the Elbow Lake region are also thought to have a similar relation to flow direction, although striations

1 to 2 miles north bear almost directly south. This divergence is attributed to the local topography, with Pine Mountain lying between the ridges and the striations, and 6 miles southeast are Rainy lobe striations trending southeasterly, nearly parallel to the ridges.

"Drumlinized" landscapes in Canada (Armstrong and Tipper, 1948, p. 287-293; Lang, Bostock, and Fortier, 1947) are striking similar, so the hypothesis is advanced that the Cook County features are primarily ridges of till deposited beneath the Rainy lobe as it moved in a southeasterly direction. Lack of deep exposures requires that this remains a hypothesis rather than a conclusion because somewhat similar features have been carved in bedrock by deep glacial scouring (Smith, 1948, p. 508-513). The term drumloid ridge is used because these features lack the asymmetrical longitudinal profile and stream-lined form usually attributed to ideal drumlins.

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ANHYDRITE-GYPSUM EQUILIBRIUM RELATIONS*

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ABSTRACT. Thermodynamic calculations show that gypsum will break down to anhydrite plus water below 14° C. in the presence of a saturated NaCl solution. Gypsum will begin to precipitate out of a concentrated sea water solution at temperatures below 34° C. Anhydrite is the first calcium sulphate to precipitate if the temperature is above 34° C. Pressure acting equally on all phases raises the temperature of dehydration. A mean pressure acting on the solid phases 2.4 times as great as that acting on the liquid phases lowers the dehydration temperature. The depth to which gypsum will be found in nature will depend on the temperature gradient, composition of groundwaters and ratio of lithostatic to hydrostatic pressure acting on the calcium sulphate deposit.

INTRODUCTION

QUANTITATIVE consideration of the anhydrite-gypsum equilibrium relations can lead to a better understanding of the factors controlling the deposition of these minerals and the alteration of one to the other. Solubility measurements (Posnjak, 1938) have shown that above 42° C. in a saturated solution of $CaSO_4$, anhydrite is the stable form.

There is, however, little experimental information as to the effect of concentrated salt solutions on the gypsum-anhydrite transition temperatures, and there is no information as to the effect of pressure on the transition. A careful study of the thermodynamic properties of gypsum and its dehydration products has been made by Kelley, Southard and Anderson (1941). Their data make possible the calculation of the effect of pressure and salt solutions on the transition temperature.

The study of this dehydration illustrates the numerical magnitude of factors that may be of considerable importance in the breakdown of hydrous minerals. It is possible to calculate with a considerable degree of accuracy the effect of different pressures on the liquid and solid phases.

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$CaSO_4$—H_2O SYSTEM

Kelley et al. (1941, p. 69) found six solid phases in the $CaSO_4$—H_2O system with distinct thermodynamic properties. These are: $CaSO_4 \cdot 2\ H_2O$ (gypsum), two polymorphs of $CaSO_4 \cdot \frac{1}{2}\ H_2O$, two polymorphs of soluble $CaSO_4$, and one insoluble $CaSO_4$ (anhydrite).[1] A plot of Kelley's free energies of the dehydration reactions shows that anhydrite and gypsum are the two most stable phases in the system (compare curve B-B with other curves in figure 1). The other forms appear in the commercial production of plaster of Paris and in experimental work in the system.

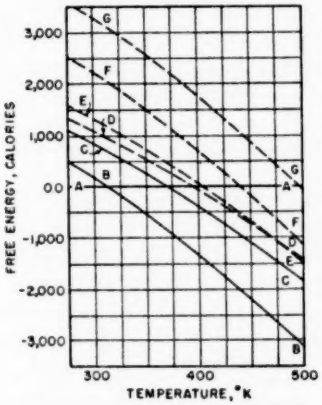

Fig. 1. Free energies of reaction (after Kelley et al., 1941).

Curve A-A	$\Delta G = 0$ along this curve.	
Curve B-B	$CaSO_4 \cdot 2\ H_2O = CaSO_4$ (insol.) $+ 2\ H_2O$	(1).
Curve C-C	$CaSO_4 \cdot 2\ H_2O = CaSO_4 \cdot 1/2\ H_2O$ (α^*) $+ 3/2\ H_2O$	(1).
Curve D-D	$CaSO_4 \cdot 2\ H_2O = CaSO_4 \cdot 1/2\ H_2O$ (α) $+ 3/2\ H_2O$	(1).
Curve E-E	$CaSO_4 \cdot 2\ H_2O = CaSO_4 \cdot 1/2\ H_2O$ (β) $+ 3/2\ H_2O$	(1).
Curve F-F	$CaSO_4 \cdot 2\ H_2O = CaSO_4$ (sol. α) $+ 2\ H_2O$	(1).
Curve G-G	$CaSO_4 \cdot 2\ H_2O = CaSO_4$ (sol. β) $+ 2\ H_2O$	(1).

[1] There is some uncertainty as to the nature of the hemihydrates and soluble anhydrites. Thermodynamic measurements by Kelley et al. (1941) show that the solid phases mentioned above have distinct thermodynamic properties. The α and β forms of the hemihydrates give almost identical X-ray patterns. Analysis of the hemihydrates shows a variable amount of water. These facts have led some workers to postulate that the soluble anhydrites and hemihydrates are structurally identical but have varying amounts of zeolitic water. If this is so, it is quite remarkable that the various forms give such consistent physical chemical data. The literature on the subject is summarized by Kelley and by Kruis and Späth (1951).

Soluble $CaSO_4$ and $CaSO_4 \cdot \frac{1}{2} H_2O$ are rare in nature. Professor Frondel pointed out to the author that the mineral bassanite may have the composition $CaSO_4 \cdot \frac{1}{2} H_2O$. In all major deposits of $CaSO_4$, however, we are dealing with the relatively most stable solid phases in the $CaSO_4$—H_2O system. If equilibrium is maintained, gypsum will break down directly into anhydrite and water. There will be no intermediate hemihydrate formed, since the hemihydrates are unstable with respect to gypsum and anhydrite plus water (see fig. 1).

BREAKDOWN OF GYPSUM TO ANHYDRITE PLUS WATER

Kelley et al. (1941, p. 44) give equation (1) as the free energy for reaction $CaSO_4 \cdot 2\ H_2O = CaSO_{4\ \text{anhydrite}} + 2\ H_2O_{\text{liquid}}$. This equation is based on the calorimetric determination of the heat of hydration of anhydrite, heat capacity measurements, and vapor pressure measurements.

$$(1) \quad \Delta G = 163.89\ T + .0215\ T^2 - 65.17\ T \log_{10} T - 2495$$

where ΔG is free energy of reaction in calories per mole and T is the temperature in degrees Kelvin.

For $\Delta G = 0$ the above equation gives 40° C. as the temperature at which gypsum, anhydrite and pure liquid water can coexist at one atmosphere total pressure (Kelley et al., p. 44, 1941). Below 40° C. gypsum is the stable phase, and above 40° anhydrite in the presence of water is the most stable form. However, 40° C. is the transition temperature only when anhydrite is in the presence of pure liquid water and when the total pressure on the system is one atmosphere.

EFFECT OF PRESSURE ON THE REACTION

An increase of total pressure on the system will favor the assemblage of lesser volume. When the pressure is acting equally on all phases of a system, the pressure dependence of the free energy of reaction is given by the following equation:

$$(2) \quad \frac{\partial \Delta G}{\partial P} = \Delta V$$

ΔV for the gypsum-anhydrite reaction is 6.66 cc per mole of anhydrite. An increase of total pressure will favor the formation of gypsum.

Employing the relations

$$(3) \qquad d\Delta G = \left(\frac{\partial \Delta G}{\partial P}\right)_T dP + \left(\frac{\partial \Delta G}{\partial T}\right)_P dT$$

$$(4) \qquad d\Delta G = -\Delta S\, dT + \Delta V\, dP$$

and equation (1), setting $d\Delta G = 0$, we obtain

$$(5) \qquad \frac{dP}{dT} = \frac{-135.59 - .043\ T + 65.17\ \log_{10} T}{\Delta V}$$

ΔV is assumed to be constant over the temperature interval 0-50° C. and the pressure interval 1 to 500 bars. This equation could easily be integrated, giving an analytic relation between the transition temperature and pressure. For the present purpose it is sufficient to obtain the initial slope $\left.\frac{dP}{dT}\right|_{40°\ C.}$ and consider the transition curve as a straight line.

We find that

$$(6) \qquad \left.\frac{dP}{dT}\right|_{40°\ C.} = 85.4 \text{ bars/deg.}$$

The transition temperature at 500 bars is 46° C.

Neglecting the effect of dissolved solids on the free energy of the water, the case considered above would correspond to the conditions prevailing during deposition of anhydrite or gypsum, since during deposition from a body of water the same total pressure would act on both the solid and liquid phases. However, this is not applicable once the gypsum and anhydrite are buried, if the water has free access to the surface. In this case, the water pressure should be approximated very closely by the hydrostatic pressure of water whose height is equal to the depth of burial. Measurements of water pressure in oil reservoirs show that for depths down to about 4000 feet the water pressure is equal to the calculated hydrostatic pressure (Cannon and Craze, 1938; Dickinson, 1953).

Assuming that the mean density of the rocks overlying the deposit of gypsum and anhydrite is 2.4, the mean pressure acting on the solid phases would be 2.4 times the pressure acting on the liquid phase. The slope of the curve in such a case

$$(7) \qquad \frac{dP}{dT} = \frac{\Delta S}{\Delta V_s + V_{H_2O}\ /2.4}$$

where ΔV_s is the difference in volume of one mole of anhydrite

and gypsum, V_{H_2O} is the volume of 2 moles of pure H_2O, and P is the pressure on the solid phases.

$$\Delta V_s = -29.48 \text{ cc per mole of anhydrite}$$

$$V_{H_2O}/2.4 = 15.06 \text{ cc per two moles of water}$$

$$(8) \qquad \left.\frac{dP}{dT}\right|_{40^\circ \text{ C.}} = -39.45 \text{ bars/deg.}$$

A pressure on the solid phase 2.4 times the pressure on the liquid phase lowers the dehydration temperature of gypsum one degree for 39.45 bars. At a pressure of 500 bars the dehydration temperature is 27° C.

The curves A and B of figure 2 show the relation of dehydration temperature to pressure; curve A represents a case where the same pressure is acting on all phases, curve B represents the case of a lithostatic pressure acting on the solid phases and a hydrostatic pressure acting on the liquid phase.

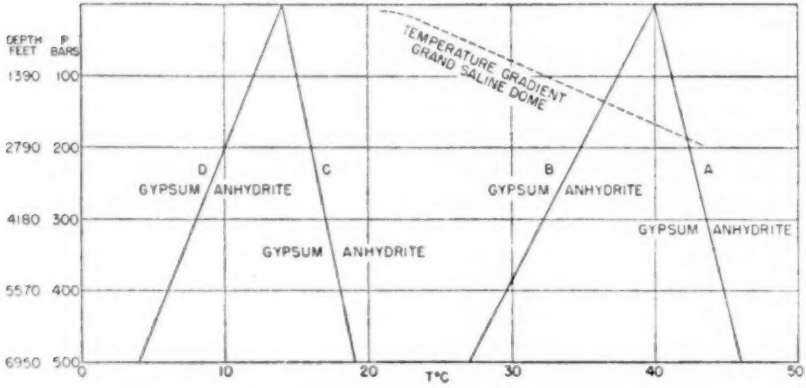

Fig. 2. Pressure-temperature relations for the reaction gypsum = anhydrite + water.

Curve A — Same pressure acting on all phases, in presence of pure water

Curve B — Rock pressure acting on solid phases, hydrostatic pressure acting on pure water

Curve C — Same pressure acting on all phases in presence of saturated NaCl solution

Curve D — Rock pressure acting on solid phases, hydrostatic pressure acting on saturated NaCl solution

EFFECT OF DISSOLVED SOLIDS ON THE REACTION

In the above calculations it is assumed that the water in equilibrium with gypsum and anhydrite is pure water. Let us consider the effect of dissolved solids on the free energy of water. The assumption is made that the water vapor in equilibrium with water can be treated as an ideal gas (a reasonable assumption for vapor pressures less than 1 bar at temperatures above 0° C.).

Salts dissolved in the water will lower the vapor pressure of the water. This lowering of vapor pressure can be related to the change in free energy of the vapor by

$$\Delta G = RT \ln p/p_o \tag{9}$$

ΔG is the change in free energy of the vapor in going from a vapor pressure p_o to a vapor pressure p. Since the water in solution will be in equilibrium with the vapor above it, ΔG also represents the change in free energy of the liquid water. Therefore, for the reaction

$$H_2O = H_2O \text{ in salt solution}$$

$\Delta G = RT \ln p/p_o$ where p_o is vapor pressure of pure water at temperature T and p is the vapor pressure of water vapor over the salt solution at the same temperature. Considering the reaction $CaSO_4 \cdot 2\,H_2O = CaSO_4$ anhydrite $+ 2\,H_2O$ in salt solution and equations 1 and 9, ΔG is given by

$$\Delta G = 163.89\,T + .0215\,T^2 - 65.17\,T \log_{10} T - 2495 + RT\ 2.303 \log_{10} p/p_o \tag{10}$$

Complete data on the lowering of vapor pressures by various salts as a function of temperature and concentration are given in the *International Critical Tables* (1933). Sodium chloride is the most abundant salt in sea water and probably the major constituent of solutions that might come into contact with buried gypsum-anhydrite deposits. The lowering of the vapor pressure of the water by natural salts is approximated by the lowering of vapor pressure by sodium chloride. Representative values of p/p_o and of ΔG for solutions of sodium chloride of various concentrations and temperatures are given in table 1.[2]

[2] Table 1 computed from data given in *International Critical Tables*.

TABLE 1

Temperature °C.	Molality of Solution	Chlorinity in Parts/Thousand	p/p_o	ΔG in Calories
20	6.136*	160	.7521	−166
30	6.168*	161	.7500	−174
40	6.207*	162	.7250	−177
20	4.967	136	.8120	−121
30	4.967	136	.8113	−126
40	4.967	136	.8116	−130
20	1.901	61	.9359	− 39
30	1.901	61	.9363	− 40
40	1.901	61	.9350	− 42

* Saturated solution

Data given in part in table 1, together with equation (10), determine the transition temperature as a function of concentration of sodium chloride. This dependence is shown in figure 3, and various values of the transition temperature are listed in table 2. Although the curve was drawn using data for sodium chloride, it can be used as an approximation for concentrated solutions of sea water of given chlorinity.[3]

In the presence of a solution containing 5 moles of sodium chloride, gypsum is in equilibrium with anhydrite at 21° C. At temperatures below 21° C. gypsum is the stable form. Similarly in a concentrated solution of sea water containing 115 parts per thousand of chlorine, gypsum and anhydrite are in equilibrium at 26° C.

In considering the effect of pressure on the transition in the presence of a concentrated salt solution, the proper molar volume of water in the salt solution must be used. For the same total pressure acting on all the phases;

$$\frac{dP}{dT} = \frac{\Delta S}{\Delta V}$$

Since $-\Delta S = \frac{\partial \Delta G}{\partial T}$, differentiating equation (10) results in

$$\Delta S = -135.59 - .043\,T + 65.17 \log_{10} T - 2.303\,R \log_{10} p/p_o \tag{11}$$

[3] Chlorinity, as used in this paper, is the mass of chlorine equivalent to the mass of halogens contained in one kilogram of sea water (Harvey, 1945). Sea water has a chlorinity of about 19 parts per thousand parts of solution.

Adams (1931) gives the partial molar volume of water in a sodium chloride solution at 25° C. for various concentrations of sodium chloride. These partial molar volumes are given in table 2. Using these data, assuming ΔV is constant over the temperature-pressure interval considered and using equation (11), dP/dT is determined for various concentrations of salt solutions. The values of dP/dT obtained are listed in table 2.

TABLE 2

Moles NaCl in Solution	Partial Molar Volume of Water (at 25° C.)	Dehydration Temperature at one bar	dP/dT in bars/deg.
.90	18.05	37	85.8
1.90	18.02	34	86.5
3.02	17.97	30	87.7
4.28	17.92	25	88.9
5.70	17.86	17	90.3
6.15*	17.84**	14	90.8

* Saturated

** Extrapolation

In a concentrated sodium chloride solution the transition curves have a greater initial slope than for pure water.

The effect of differential pressure on the dehydration reaction in the presence of salt solution can now be calculated. A rock density of 2.4 is assumed. The density of the liquid is assumed to be that of a sodium chloride solution of the given concentration. The slopes of the dehydration curves for differential pressures are given in table 3.

TABLE 3

Moles NaCl in Solution	Density of Solution	dP/dT in bars/deg.
.90	1.03	−40.6
1.90	1.07	−42.4
3.02	1.11	−44.1
4.28	1.15	−45.9
5.70	1.19	−47.7
6.12*	1.20	−48.3

* Saturated solution

The relation of temperature of dehydration to pressure in the presence of a saturated sodium chloride solution is shown in figure 2. Curve C represents the case of a differential pressure.

The transition temperature of gypsum to anhydrite at one bar pressure as a function of salt concentration is given in figure 3. The effect of pressure on the transition in the presence of salt solutions is given by values of dP/dT listed in tables 2 and 3.

By using these data it is possible to calculate a set of curves similar to A, B, C, and D of figure 2. The curve defining the transition in the presence of a 1.9 molar solution of NaCl has an initial point at 34° C. (fig. 3). The slope of the curve for the case of different pressures acting on the liquid and solid phases is −42.4 bars/degree (see table 3). Similarly, transition curves for other salt concentrations can be calculated.

COMPARISON OF CALCULATION WITH EXPERIMENT

On the basis of solubility measurements Posnjak (1938) determined the temperature of dehydration of gypsum to anhydrite to be 42° C. at one atmosphere pressure. This value is 2° higher than the value calculated using thermodynamic

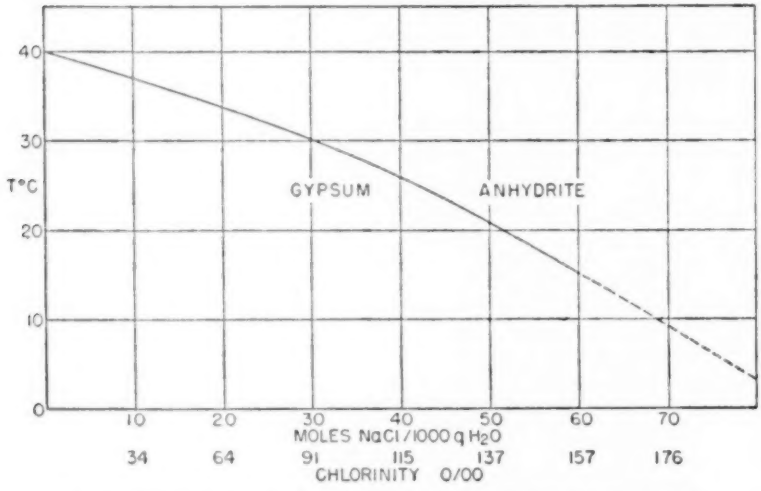

Fig. 3. Dependence of the dehydration temperature of gypsum on concentration of NaCl in solution at one bar pressure.

data. Posnjak (1940) determined the solubilities at 30° C. of anhydrite and gypsum in a salt solution of essentially the composition of concentrated sea water. The transition of anhydrite to gypsum took place at 30° C. in a solution containing 93 parts per thousand of chlorine. The calculated temperature for a solution of this chlorinity is 30° C. (fig. 3). There is no experimental information as to the effect of pressure on the dehydration temperature.

GEOLOGIC CONSIDERATIONS

Deposition.—About 1.26 grams of calcium sulphate are present in 1000 grams of natural sea water; this amount is less than the solubility of calcium sulphate in water. The total salinity of sea water has to be increased by a factor of 3.35 to obtain saturation with respect to calcium sulphate (Posnjak, 1940). This salinity corresponds to a chlorinity of about 65°/₀₀. Assuming that the above solubility relations, obtained at 30° C., hold over the temperature interval considered, figure 3 shows that in a solution of this chlorinity the transition temperature is 6° less than in pure water. Therefore gypsum will precipitate out of sea water at all temperatures below 34° C.[4] if the deposition is an equilibrium process. The value often quoted in the literature is 42° (Phillips, 1947) based on Posnjak's work. This value refers to a solution saturated with $CaSO_4$ but containing no other salts. The presence of other salts in the proportion found in natural sea water will lower the transition temperature by about 6°. Therefore, anhydrite will be the phase precipitated out of sea water if the temperature is above about 34°. Pressure has little effect on this temperature; a pressure of 100 bars will raise the temperature to 35° C.

The history of crystallization of calcium sulphates from a body of sea water can be traced if a sea water temperature is assumed. At 25° C. salts will be concentrated by evaporation until a chlorinity of 65°/₀₀ is reached. At this point gypsum will begin to precipitate. Gypsum will continue to precipitate until the chlorinity reaches a value of 119°/₀₀; then anhydrite will precipitate. Gypsum in contact with sea water at chlorin-

[4] Here, and in the discussion which follows, temperatures derived from thermodynamic data will be used. As mentioned previously, these temperatures are within 2° of experimentally determined dehydration temperatures.

ity above 113°/$_{oo}$ will be unstable and will break down to anhydrite.

In saturated NaCl solutions gypsum will precipitate only at temperatures below 14° C. as is shown in table 2. In concentrated sea water gypsum and halite will precipitate together at a chlorinity of 184°/$_{oo}$ and at temperatures below 7° C. The presence of salts other than NaCl in concentrated sea water lowers the maximum temperature at which gypsum and halite can coexist from 14° C. to 7° C.[5]

Relation of gypsum to depth.—A knowledge of the temperature gradient in a region of calcium sulphate deposits and the data calculated above determines a maximum depth at which gypsum will be stable. There probably is a transition region over which both gypsum and anhydrite will appear. Below this region no gypsum will be found.

Deposits of calcium sulphate are commonly found with salt domes. Since temperature gradients over salt domes are available, a comparison of the predicted depth to which gypsum should be found and observed depths is possible for calcium sulphate deposits associated with salt domes.

Hawtoff's (1930) gradient over the Grand Saline Dome in Van Zandt County, Texas, is plotted in figure 2. A mean rock density of 2.4 has been assumed. This temperature-depth curve intersects the dehydration curve for a differential pressure acting on the solid and liquid phases at a depth of 1820 feet. For this temperature gradient, gypsum would not be in equilibrium with pore water below about 1820 feet.

The depth to which gypsum will be found is dependent on the amount of dissolved solids in the ground water. The dehydration curve for gypsum in the presence of a solution of chlorinity 61°/$_{oo}$[6] will intersect the temperature-depth curve at about 1300 feet. Goldman (1952, p. 61) notes that the maximum depth to which gypsum is found in cap rock is about 2000 feet. In the Sulfur Salt Dome, which Goldman studied in detail, gypsum was rarely found below 1180 feet. Considering

[5] This calculation assumes that salts other than NaCl have no effect on the solubility relations of NaCl. Therefore, 7° C. may be somewhat low for the simultaneous precipitation of halite and gypsum from concentrated sea water.

[6] Timm and Maricelli (1953) have studied the salinities of formation waters in southwest Louisiana. Salinities up to 4½ times that of natural sea water have been found.

the uncertainty in temperature gradients and groundwater composition, the agreement between calculated and observed depths of disappearance of gypsum is satisfactory.

Curves C and D of figure 3 represent the conditions necessary for the breakdown of gypsum to anhydrite in the presence of a saturated salt solution. Pore water in a halite bed would be saturated with NaCl. Gypsum would not be in equilibrium within this pore water except at temperatures below 14° C. Gypsum is not in equilibrium with buried halite except at temperatures lower than 14° C. Published cross sections of cap rock (Taylor, 1938) show a layer of anhydrite separating the gypsum from the rock salt. This sequence is interpreted in the following way: The pore water near the main body of salt probably has a high concentration of NaCl. Therefore, with any normal temperature gradient, anhydrite would be the phase in equilibrium with the pore water. At a higher level the groundwater will no longer be saturated with NaCl, the temperature will be lower, and gypsum will be the stable calcium sulphate.

Salt deposits.—Major deposits of calcium sulphate are found associated with undisturbed salt beds. A recent detailed study of the Permian salt deposits in Yorkshire by Stewart (1949, 1951a, 1951b) permits a comparison of the sequence of deposition derived from the present paper and that observed in a natural deposit.

Stewart describes three evaporite beds, each showing the general sequence (from bottom to top): Anhydrite-carbonate rocks, anhydrite-halite-polyhalite rocks, and rocks consisting dominantly of chlorides. Gypsum was not found in any of the core samples. The absence of gypsum can be explained by the fact that the beds are now at a depth of 3600 feet to 5000 feet. The temperature at 4200 feet is 42° C.; at 3000 feet it is 37° C. (Lees and Tait, 1946, p. 269). Therefore the maximum depth at which gypsum would be stable in the presence of pure water would be about 2500 feet, assuming different pressures acting on the liquid and solid phase (fig. 2). In his earlier paper Stewart (1949, p. 659) stated that the basal anhydrite rock was deposited as primary anhydrite. However, in a recent paper (1953) he reports on evidence that the lower anhydrite-carbonate rocks contain anhydrite pseudomorphs after gypsum. The intermediate zone of the evaporite beds contains

abundant anhydrite pseudomorphs after gypsum; these pseudomorphs are associated with halite. Complexly interbedded with the rocks containing pseudomorphs are rocks made up of primary anhydrite and halite. Stewart (1949, p. 659) interprets the sequence as the result of alternate deposition of gypsum-halite and anhydrite-halite. With this information it is possible to correlate Stewart's observation with material presented in this paper.

On evaporation gypsum is the first calcium sulphate deposited. Continuing evaporation concentrates other salts and eventually anhydrite is deposited instead of gypsum. The concentrated brines are no longer in equilibrium with gypsum and would alter previously deposited gypsum to anhydrite. There then follows a period of alternate deposition of anhydrite-halite and gypsum-halite. Anhydrite-halite rocks were deposited when the temperature of the bottom waters was above about 7° C., and gypsum-halite rocks were deposited when the temperature was below 7° C.

The low bottom water temperatures indicated by the original gypsum-halite assemblage give rise to several possibilities as to the climate of the region. The evaporating basin could have been in a cold but dry area, the temperature of the basin varying only a few degrees about 7° C. Alternatively, the basin could have been in an area of large fluctuations in temperature, the gypsum-halite beds representing warmer periods.

Perhaps the major objection to this picture is the low temperature required for deposition. The treatment presented assumes the proportions of salts in the Permian seas was essentially that of the present time. Decreasing the concentrations of salts other than NaCl and $CaSO_4$ increases the temperature at which gypsum and halite could form simultaneously, but in no case could gypsum and halite be in equilibrium with water above 14° C.

Another possibility is that gypsum and halite were not deposited at the same time. A history of crystallization compatible with thermodynamic data is as follows: Gypsum is deposited, evaporation leads to deposition of anhydrite, the brines in equilibrium with anhydrite replace the previously formed gypsum. Deposition of halite and anhydrite follows. An influx of fresh water into the basin reduces the concentration of salts, halite is dissolved, and anhydrite is converted

to gypsum. Continuing evaporation results in brines capable of first replacing gypsum with anhydrite and then with anhydrite and halite.

Such a cycle could be repeated several times, leading to a complex interlayering of anhydrite-halite rock and rocks containing pseudomorphs after gypsum. The above sequence could form in water having a temperature lower than 34° C.

SUMMARY

An attempt has been made to correlate available thermodynamic, experimental, and geologic information on the system $CaSO_4$—H_2O.

Using the thermodynamic data, the temperature of breakdown of gypsum to anhydrite plus water as a function of the amount of salt dissolved in the water, has been calculated. High salt contents lower the dehydration temperature.

The effect of pressure on the transition temperature has been calculated. An increase of pressure acting equally on all the phases raises the dehydration temperature of gypsum. On burial of a calcium sulphate deposit the pressure acting on the solid phases is probably unequal to that on the liquid phases. Under these conditions the temperature of transition is lowered.

The maximum depth to which gypsum will be found is a function of the temperature gradient over a region, the composition of pore water, and the ratio of hydrostatic to lithostatic pressure. Using the gradient found over the Grand Saline Dome in Texas, assuming the ground water to contain 10 percent dissolved salts, the maximum depth at which gypsum is stable would be 1360 feet in this locality. Lower temperature gradients would increase the depth to which gypsum is found, as would lower concentrations of salts in the ground water.

Stewart describes rocks that represent alternate deposition of anhydrite-halite and gypsum halite. In light of material presented in this paper this means that the bottom waters in the basin were around 7° C. during the deposition of the halite-gypsum rocks.

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MEAN AND STANDARD DEVIATION OF CERTAIN GEOLOGIC DATA — A DISCUSSION

Zdenek V. Jizba

ABSTRACT. The applicability of standard statistical equations to some geologic problems is questioned. Examples show that when an origin for data can be arbitrarily chosen the mean and standard deviations are invalid. In those situations where no natural base line is available and the origin is arbitrarily picked, two investigators may arrive at different conclusions using the same data and equations.

INTRODUCTION

EVEN the most casual perusal of the geologic literature indicates a recent increase in the use of statistics in the attempt to achieve a better understanding of geologic processes and earth history. This is an altogether commendable and healthy trend which holds great promise of giving new and quantitative precision to our studies and of placing in our hands a powerful tool with which to attack many problems still posed by the rock record and by the processes involved in creating that record.

It is also apparent that the increased use of statistics meets with varied reactions from different investigators. Those who are unfamiliar with the techniques and potentialities of statistics are apt to pass off any statistical treatment with a scornful "but you can prove anything with figures." Then there are others who feel that a report without "figures" is not "sound" and, feeling thus, they tend to adorn their manuscripts with "figures" even though they may add nothing to the value of the study. A third group is represented by those who realize the value of a statistical analysis and use their data in a sincere and honest effort to achieve a better understanding of their problem and to establish facts and relationships that would not be demonstrable without the aid of statistical analysis.

An excellent paper on the orientation of quartz grains in the Bradford Sand written by Griffiths and Rosenfeld and recently published in this Journal (1953, Vol. 251, p. 192-214) represents the type of work being done by members within this third group. Because of the high quality of this report, there-

fore, it is with no little humility that this writer cites the appendix to this article as containing an example of a statistical method which could be incorrectly applied in geologic problems.

THE METHOD OF MOMENTS

Griffiths and Rosenfeld devote the appendix of their paper to the discussion of a randomness criterion obtained by applying rectangular frequency distribution to the problem of angular distributions. This method can be found in most introductory books on statistics.[1] The method is used to determine mean and standard deviations which in turn are measures of randomness or preferred orientation. The writer feels that the method as described by Griffiths and Rosenfeld is limited in its application to many geologic problems, including certain analyses of rock fabrics. The technique is usable only in those distributions where all the points are concentrated in one local area of the projecting sphere or where the origin is defined. In the main body of their paper Griffiths and Rosenfeld correctly apply the technique, but neither there nor in the appendix is there any specific discussion of its limitations as mentioned above. For this reason it seems proper to point out clearly the conditions under which the method may be used to determine random and preferred orientation as measured by mean and standard deviations.

The method of moments cannot be applied to many geologic problems demanding the determination of preferred and random orientation for two reasons:

1. The method does not define a base line, and for this reason the results obtained by one investigator are apt to be different from those reached by any other. This does not mean to say that natural base lines do not exist in geologic problems. They do, and as an example we can cite the horizontal plane in the study of sedimentary rocks and the "S" surfaces in metamorphic rocks. In each of these situations it would be just as wrong to overlook this naturally defined origin, base line or reference plane, as it is wrong to create an artificial base line where one is not present.

[1] The last equation on p. 211 should read $K = 1/2a$ instead of $K = \frac{1}{2}a$.

2. Furthermore, the result does not necessarily imply random or preferred orientation.[2]

An examination of the equations contained in the appendix to the paper by Griffiths and Rosenfeld shows that both the mean and the standard deviation have no validity whenever the origin may be chosen arbitrarily and thus shifted at the discretion of the investigator.

For example, assume a variable, x, in degrees and a frequency distribution which exists in the whole range 0°-360° as shown in figure 1. This frequency distribution may or may not show a peak. If the origin is taken at 0°, then the mean ($\bar{x}$) is given by:

$$\bar{x}_A = \frac{\Sigma\, x_j}{n}$$

Now shift the origin by an angle a such that at least one point xj will be moved from one end of the distribution histogram to the other. Histograms A and B of figure 1 represent two different interpretations of the same data. In A the origin is chosen at 0° which gives a mean at 98.33° and a standard deviation of 36.41°. In B the origin was chosen at 0° + a which gives a mean at 105.00° and a standard deviation of 36.51°. The mean and standard deviation for the two histograms are not the same because the value xj has been shifted from one end of the histogram to the other. The mean x_1 of histogram B in terms of the mean of histogram A is given by:

$$\bar{x}_B = \frac{\Sigma\, x_j}{n} + \frac{180}{n}$$

From this argument it can be seen that unless there is an adequate method of defining the origin, it is entirely possible

[2] The question of random orientation has been much discussed in geology; e.g. Krumbein (1950), (1935); Chayes (1950); Pincus (1952); and Rodgers (1952). Although there may not be a complete agreement on a definition of random orientation, it can be said in general that it should have the following attributes:

a) Whatever the space R being considered (e.g. ellipse, circle, the surface of a sphere, or an ellipsoid), a random distribution is such that the expected value of the density is the same for any subregion r in this space.

b) The distribution of points in this space should remain random as defined above for any permissible change of the reference plane, origin, or base line.

for two investigators to arrive at different values for the mean.

Similarly, the standard deviation depends on the choice of

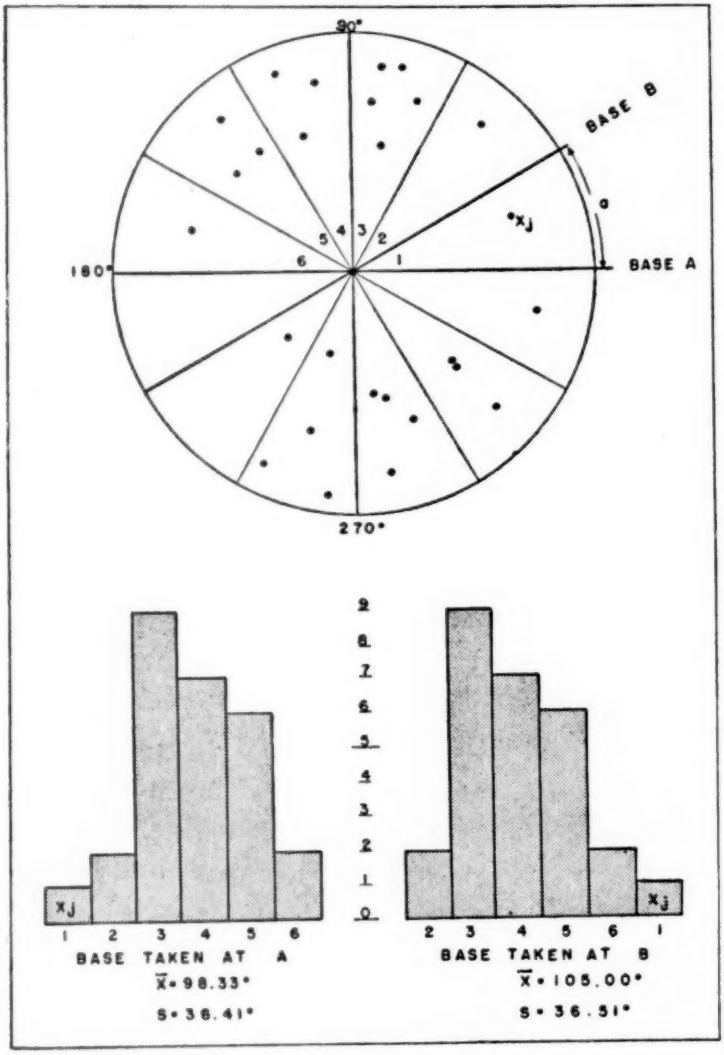

Figure 1. Top: distribution of points plotted on a stereographic projection. Bottom: histograms A and B showing the distribution of points from the stereographic projection.

the origin. The variance s^2 (standard deviation squared) is given by:

$$S^2_A = \frac{\Sigma (x_i - \bar{x})^2}{n}$$

With the change of origin by the angle a, the new variance S^2_1, will be given by:

$$S^2_B = \frac{\Sigma \left(x_i - \bar{x} + \frac{180}{n}\right)^2}{n} + \frac{1}{n}\left\{(x_j + 180) - \left(\bar{x} + \frac{180}{n}\right)\right\}^2$$

which reduces to:

$$S^2_B = S^2_A + h^2$$

and where:

$$h^2 = \left\{\left(\frac{180}{n}\right)^2 (n^2 - n + 1) - 360\ (\bar{x} - x_j)\right\}$$

AN EXAMPLE

The danger inherent in the use of the method of moments in the determination of mean and standard deviation as discussed above can be illustrated with an example based on data given by Krumbein (1939). The example chosen is confined to Krumbein's use of the strike of the elongation of glacial outwash pebbles and not to his use of dips. In this study (p. 699-702) the azimuths of the *a*-axes of 100 pebbles were measured in the field and these data were then analyzed by the method of moments. Taking the origin at 90°, Krumbein found the mean to be 359.6° and the standard deviation 52.4° (table 1).

The origin was chosen at 90° apparently because the field observations indicated that the waters which deposited the outwash flowed south or somewhat east of south. The field observation is not questioned, but the use of this direction as a base line (and hence as a determinant of a point of origin) is not valid if the analysis of the data is designed to demonstrate the direction of stream flow. For instance, assume that no field observations were available and we had only the statistical data on hand to determine the flow direction. Under such

TABLE 1

Interval	Frequency f	Class Interval u	fu	u^2	fu^2
90-110	14	−4	−56	16	224
110-130	12	−3	−36	9	108
130-150	6	−2	−12	4	24
150-170	7	−1	− 7	1	7
170-190	16	0	..	0	0
190-210	13	1	13	1	13
210-230	9	2	18	4	36
230-250	14	3	42	9	126
250-270	9	4	36	16	144
Sum	100	0	− 2	xx	682

Frequency table from figure 13, p. 699, Krumbein (1939). Origin taken at 90°.

a condition we would have no indication as to a point of origin and could arbitrarily pick any point. Suppose that, using the same data as provided by Krumbein, we were to pick a point of origin at 150° instead of 90° (table 2). The mean deviation would be 207.2° and the standard deviation 48.4°.

Griffiths and Rosenfeld list the standard deviation for a random distribution as 51.96°. If we apply this value to Krumbein's original calculation, the distribution is found to be essentially random as inferred by him. On the other hand, using the same data but changing the origin from 90° to 150°,

TABLE 2

Interval	Frequency f	Class Interval u	fu	u^2	fu^2
150-170	7	−4	−28	16	112
170-190	16	−3	−48	9	144
190-210	13	−2	−26	4	52
210-230	9	−1	− 9	1	9
230-250	14	0	0	0	0
250-270	9	1	9	1	9
270-290	14	2	28	4	56
290-310	12	3	36	9	108
310-330	6	4	24	16	96
Sum	100	0	−14	xx	586

Frequency table of the same data as table 1, but the origin taken at 150°.

as suggested above, we find that the distribution is not so conclusively random. Actually in these two results the smaller standard deviation produced by the second choice of origin (150°) would indicate, *if anything*, that this second choice is better than the one used by Krumbein (90°).

SUMMARY

The method of moments is of use only when a base line can be defined. Where no such natural base line exists, the statistical analysis is neither reproducible nor can we say that any single analysis based on any particular base line is valid.

With no intent to criticize the pioneering work of Krumbein nor the analyses of Griffiths and Rosenfeld, the writer feels that many of the difficulties encountered in the application of statistics to geologic data could be eliminated if two points were remembered:

1. Statistical methods of analysis do not *prove* an argument. A statistical approach, however, determines whether an assumption is true or false within certain "confidence limits." This generally means that the probability of accepting an erroneous conclusion is small ("error of the first kind") and that the probability of rejecting the same conclusion if valid is at a minimum ("error of the second kind").

2. Statistical methods can be applied only with a full understanding of the geologic *and* the statistical problem. The equations listed in elementary textbooks of statistics may not always be applicable to many problems arising in geology. And if they do not apply, their use can often lead to erroneous conclusions in the solution of the geologic problem.

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In the preparation of this discussion the writer received both encouragement and helpful criticism from Professor Sheldon Judson, Raymond C. Murray, and Elaine Geisse, Department of Geology, University of Wisconsin. Dr. E. R. Immel of the Department of Mathematics of the University of Wisconsin has kindly read the manuscript.

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STATISTICAL DEVICES AND PETROGRAPHICAL PROBLEMS — A REPLY

J. C. Griffiths

The limitations of our test of randomness based on a rectangular frequency distribution are contained in the appendix describing the test (Griffiths and Rosenfeld, 1953) and are:

1. The operational definition (appendix paragraph 2, p. 210) of the inclination measurement which leads to a two coordinate frequency distribution (i.e. two dimensions).

2. The demand for unimodality (paragraph 2, p. 211).

Mr. Jizba's introduction of three dimensionalism leads to loss of uniqueness in our definition—a feature which commonly arises when conceptual ideas are confused with operational objectives.

Similarly the demand for a "natural base line" is confusing; the sole requirement is a definition leading to an *unique* base line which is the aim of our operational definition. As long as the definition is operationally unique, different investigators should be able to achieve the same solution. Such is not necessarily the case for a "natural base line"; e.g. bedding as normally used in geology is certainly not unique and implies different things to different people.

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REVIEW

Reports of the Swedish Deep-sea Expedition, 1947-48; HANS PETTERSSON, editor. Vol. 2, *Zoology*, fasc. 1. Göteborgs Kungl. Vetenskaps-och Vitteshets-Sämhalle. Pp. 148. Gothenburg, 1951 (Oceanografiska Institutet, Kr. 25).—This fascicle contains Nybelin, O., *Introduction and Station List;* Kullenberg, B., *On the Shape and the Length of the Cable during a Deep-Sea Trawling;* Mortensen, Th., *Echinoidea;* Gislén, T., *Crinoidea;* Silén, L., *Bryozoa;* Madsen, F. J., *Asteroidea;* Fage, L., *Sur un Pycnogonide;* Carlgren, O., *Actiniaria;* Madsen, G. J., *Ophiuroidea;* Kramp, P. L., *Hydrozoa and Scyphozoa;* Eliason, A., *Polychaeta.* The zoological work of the expedition was confined to dredging in very deep water; the results though quantitatively limited are of extraordinary interest in view of the scanty information available on life below 3000 m. The technical problems of collection are considered from a physical point of view in Kullenberg's paper which will be of immense help in planning work on other expeditions. The taxonomic work is of the high standard that is to be expected of the contributing authors. All papers give complete summaries of the distributions below 3000 m. of the groups that they consider. A few general comments may not be inappropriate. Fifteen species of determinable echinoderms are reported; in every case the species was a previously known one. The records confirm in a most satisfactory manner the conclusion of earlier investigators that this abyssal fauna is essentially a cosmopolitan one, represented, with little differentiation, in the depths of all the major oceans. In some cases there seems to be an affinity between the Indian and Atlantic rather than between the Indian and Pacific Oceans. Among the other groups 3 bryozoa and one pycnogonid are recorded, all having been named by earlier workers. The soft-bodied forms produced a few new species, but much of the material was too fragmentary for accurate determination. In the Polychaets a species of *Macellicephala,* probably *M. mirabilis,* McInt., occurred at 7625 - 7900 m. and at the time represented the maximum extension of animal life into the ocean. Subsequently, preliminary reports from the Galathea expedition indicate animal life at even greater depths.

Dr. Nybelin and his collaborators have produced a work which, reporting a small but interesting collection, will always be one of the fundamental documents of deep-sea zoogeography.

G. E. HUTCHINSON

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ERRATUM

On page 765 of this volume (October 1953) the review of Werner Nowacki's *Moderne allgemeine Mineralogie* (*Kristallographie*) contains Professor Nowacki's name incorrectly spelled in one place, and correctly in another. The reviewer and others who read proof of this review sincerely regret that this typographical error was not found in time to correct it before publication.

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